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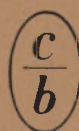
January, 1995

LITHOLOGY AND MINERAL RESOURCES

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SOME ASPECTS OF QUATERNARY SEDIMENTATION OF ORGANIC MATTER IN THE NORTHERN PART OF THE PACIFIC OCEAN

D. E. Gershanovich

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This article considers the main phases of Quaternary sedimentation of organic carbon in northern Pacific-Ocean sediments. It was established that the greatest difference between Holocene and Pleistocene sedimentation is manifested in marginal seas in the northwestern and northern part of the Pacific Ocean and south of Alaska. An important characteristic in the distribution of organic carbon is its unevenness, which is due to a change in bioproduction and hydrodynamic situations and is created by latitudinal and circumcontinental zonality. Many features of sedimentation of organic carbon depend on the bottom relief. The basic types of vertical circulation are distinguished. It is shown that vertical circulation of waters is comparable to horizontal circulation in its significance in the system of hydrodynamic factors of sedimentation of organic carbon.

Investigations of organic matter in bottom deposits in the extreme northern part of the Pacific Ocean performed in recent decades deserve special attention. This is due to the significant scale of geological and geochemical works conducted there and their close coordination with the whole set of oceanographic and hydrobiological studies of the aquatory [underwater territory] north of 30° north latitude. Moreover, the vast Northern Pacific-Ocean region, which encompasses marginal seas and gulfs of East Asia, the Bering Sea, and Alaska, island arcs, and adjacent depths of ocean basins and ridges, is distinguished by a diversity of natural conditions creating a broad spectrum of marine and oceanic sedimentation situations. The transition from the polar climatic zone to the subtropical one is quite significant within these boundaries, on dry land, as well as in the ocean, which is particularly manifested in relation to mobilization and delivery of sedimentary material, the distribution of organisms, and biological productivity, and accordingly in accumulation of organic matter on the bottom. In the sedimentary process and the course of differentiation of material in the region, an important role belongs to circumcontinental zonality and the factors determining it. Research results have revealed regions of intensive neogenesis and increased accumulation of organic matter, and regions where it is minimal. The differences are great and reach one-two orders or more (Fig. 1).

BASIC PHASES OF STUDY OF SEDIMENTATION OF ORGANIC MATTER IN THE NORTHERN PART OF THE PACIFIC OCEAN

Investigations of organic matter in the northern part of the Pacific Ocean can be divided into three phases corresponding mainly to different technical possibilities of marine expeditionary work and collection of sedimentary material. The first of these is connected with general study of the conditions and sedimentation in the northern basin of the Pacific Ocean, and with obtaining primary characteristics of the composition and distribution of bottom sediments and their content of organic matter. Naturally, the works were conducted with the samples most accessible for collection and subsequent lithological and geochemical analyses, and also for the purpose of establishing the role of plankton and benthic organisms in processes of accumulation and distribution of organic matter on the bottom. A survey of materials relating to this phase is given in the well-known publications of P. A. Bezrukov [1, 2], O. K. Bordovskii [6], A. P. Lisitsyn [20, 21], K. Emery [37], and E. A. Romankevich [28]. It was this phase, which lasted from the 1940's through the 1960's, that served as the basis for disclosure of general and regional patterns in the sedimentogenesis of organic matter over vast aquatories of the Pacific-Ocean north. Its results have been covered in great detail in [1, 2, 6, 7, 11-13, 15, 19-21, 29, 30, 37, 38, etc.]. They have also entered into many generalizations on the geochemistry of sediments' organic matter and made it possible to compile quite indicative maps of its distribution. A fundamentally important feature of the investigations during this

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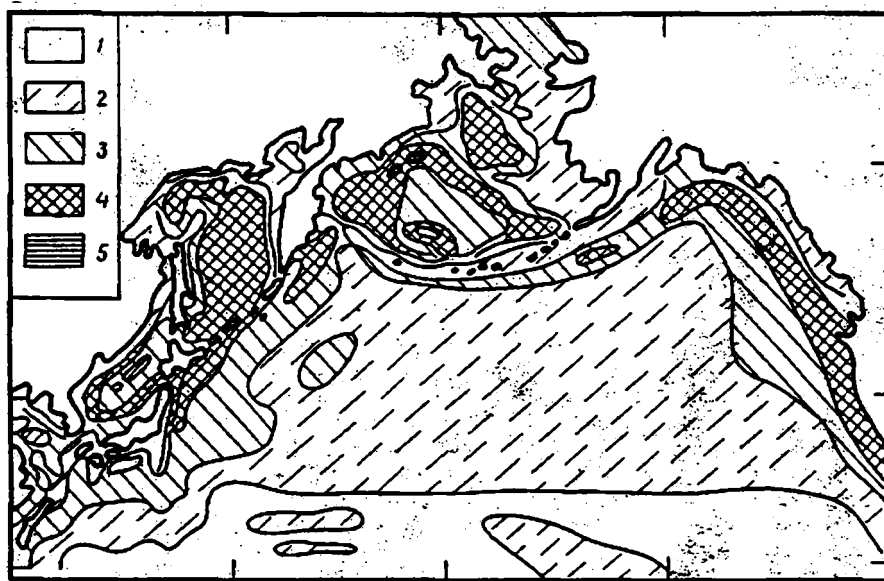


Fig. 1. Contents of organic carbon in Holocene sediments of the Northern Pacific-Ocean region [31], %: 1) <0.25; 2) 0.25-0.5; 3) 0.5-1.0; 4) 1.0-2.0; 5) >2.

phase was their direct coordination with the overall acquisition of knowledge about the course of the sedimentary process in the region and factors that determine, to the greatest degree, increased or reduced accumulation of organic matter and its distribution in specific areas.

The second phase began somewhat later (in the 1950's-1960's). In most cases, it was a direct continuation of the first phase and was a result of the possibility of using long coring tubes (as much as 10 m or more) in expeditions to collect sediment samples. In this case, the total number of samples collected also rose. The investigations encompassed new regions. With the help of these tubes, multimeter ground columns were extracted, in which, as a rule, the bedding, features of the composition and properties of surface and subsurface sediments corresponding to their sedimentation situation, and also leading traits of their stratification, lithology, and geochemistry were well preserved. Study of the columns and the organic remains contained in them showed that they encompass horizons of sediments belonging to the Holocene and Pleistocene, and, in rare cases, to the upper part of the Pliocene. A common feature of stratification seen in most columns is revealed: a change in sediments at the boundary of the Holocene and Pleistocene and, what is perhaps even more important, alternation of deposits of glacial and interglacial epochs. In many parts of the region, this indicates fairly sharp periodic changes in the whole natural situation in the northern Pacific-Ocean belt, although the degree of their expression did not remain the same in space or time at small and large depths and in regions with different rates of entry and deposition of sedimentary material. It is quite indicative that this was also traced in the sediments' lithology and geochemistry, in the accumulation of organic matter, and what is perhaps especially significant, in the composition and amount of organic remains, and their distribution on the bottom in different epochs. Of particular interest was the correspondence that was established between these features of the sedimentation process in such a vast and geologically and geochemically heterogeneous region as applied to the course of accumulation of organic matter. Materials covering this aspect of sedimentation in the northern Pacific-Ocean regions are presented in [5-7, 20, 21, 28, 29, 37, etc.]. It should be noted how important micropaleontological investigations of sediment samples and especially careful correlation of the results of diatomic analysis were in this case [16, 17, etc.]. The location of a considerable part of the Northern Pacific-Ocean region in the northern belt of silica accumulation in the World Ocean [21] precisely through remains of diatoms as the main source of organic matter arriving at the bottom determined sharp variations in its accumulation in Quaternary deposits. The role of certain other organisms in this process was also established [8, 18, 29, etc.] (Fig. 2).

In contrast to the first two, which are fairly close related to each other, the third phase of study of organic matter in northern Pacific-Ocean sediments occupies a special position. It became feasible after deep-water drilling in oceanic depths was found to be possible. In 1966, in the northern part of the Pacific Ocean, the first voyages of the drilling ship

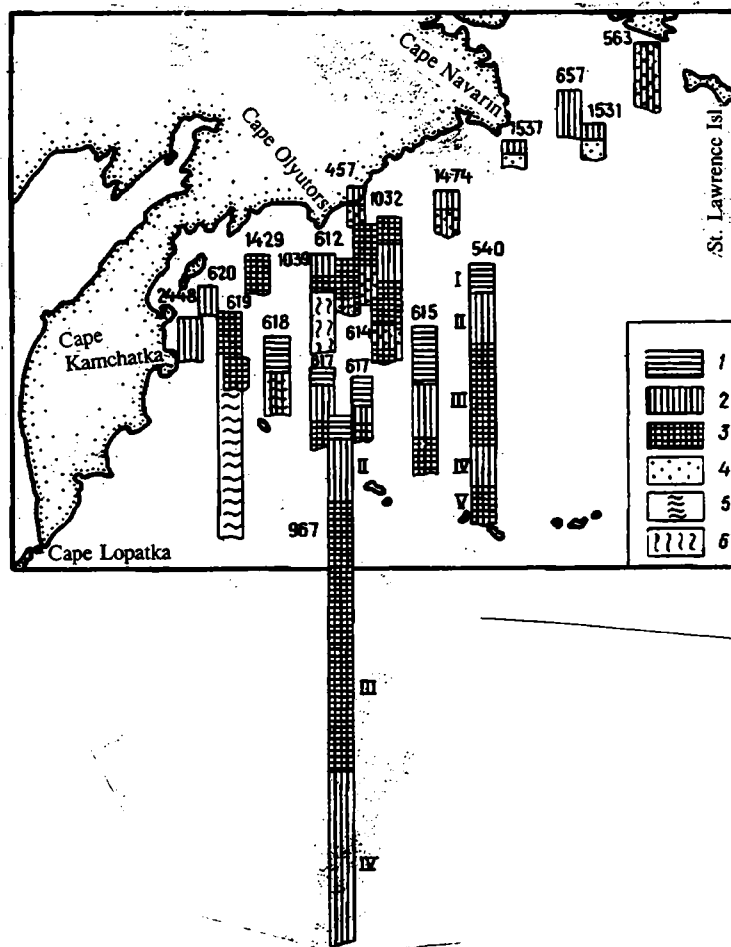


Fig. 2. Overall diagram of Quaternary stratification according to floristic complexes of diatoms in the western part of the Bering Sea [16]: 1) oceanic; 2) neritic; 3) mixed; 4) sublittoral horizon II; 5) diatoms of Tertiary age (redeposited); 6) foraminifers.

"Glomar Challenger" were carried out close to the northeastern edge of North America (voyages 5 and 6), somewhat later in the Gulf of Alaska, in the southern part of the Bering Sea, and in certain regions of the Aleutian arc (voyages 18-20), then in the northwestern part of the Pacific Ocean, along the Hawaiian and Emperor ridges, in the Sea of Japan (voyages 31, 32, etc.), and east of Japan north of 30° north latitude (voyages 56, 57, and 86). Additional material began to come in the later 1980's, when investigations were carried out on the new drilling ship "JOIDES Resolution" in the northwestern part of the Pacific Ocean, including the Sea of Japan (voyages 127 and 128). Altogether, drilling works have been carried out at more than 100 stations north of 30° north latitude, with samples of organic carbon being successfully collected at the overwhelming majority of the stations. Materials from the investigations have been published in the corresponding volumes of the drilling ships' reports.

The results of many of them were partially assembled in the monographs of É. M. Galimov and L. A. Kodina [10], and V. Ya. Trotsyuk and M. M. Marina [33] (Fig. 3), in articles on the results of drilling in sections devoted to the content and distribution of organic matter, and in the generalizations of A. P. Lisitsyn [28, etc.], E. A. Romankevich [30, 31], and others. For the northern part of the Pacific Ocean, they give data on the accumulation of organic carbon and certain other components of organic matter over the course of the Neogene and Late Paleogene; and for individual regions, for an even earlier time. A synthesis of the data is included in Yu. A. Bogdanov and M. A. Levitan's description of the history of biogenic sedimentation in the ocean [9, 18]. Thus, the total number of reports on organic matter, while not so large, still makes it possible to draw a number of interesting conclusion about how the accumulation of organic matter occurred and changed in the northern part of the Pacific Ocean, especially in the late Cenozoic.

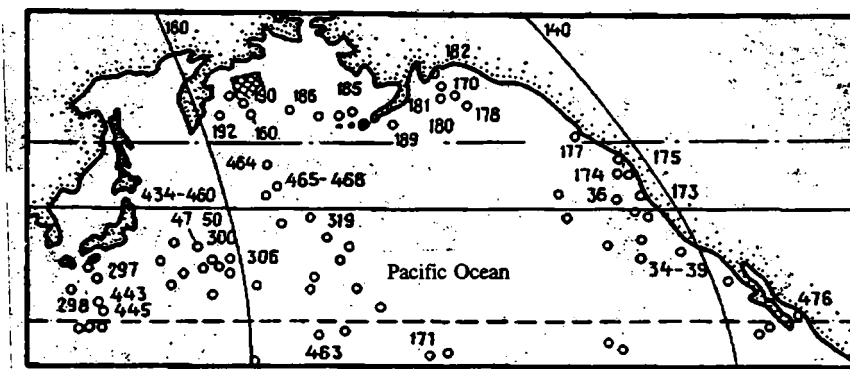


Fig. 3. Deep-water boreholes in the Northern Pacific-Ocean region [33].

In the light of all of these data, the northern part of the Pacific Ocean, with its vast marginal seas, is characterized by unevenness and significant contrast in the neogenesis, distribution, and accumulation of organic matter in sediments. The maximum variations have been established within the bounds of underwater continental margins, marginal seas, deep-water ridges, and island arcs. It must be emphasized that this is exactly where the geology and bottom relief, and the degree of its dissection, as follows from the latest research, has a great effect on the overall course of sedimentation in general and of organic matter in particular. It shows up directly through a change in depths, and in the alternation and unevenness of relief forms, which are determined by the structural inhomogeneity of the bottom and geological history. But it acts indirectly to an even greater degree, by altering the sedimentation situation and the supply of sedimentary particles, and by transforming individual sedimentation factors. The direct action frequently proves to be more localized than the indirect action, since the latter is especially closely connected with climate fluctuations, the high mobility of water masses and biocenoses, the lesser temporal constancy characteristic of them, and coverage of large areas of aquatory. One cannot fail to pay attention to the fact that the degree of contrast in accumulation of organic matter is most significant in sediments of the Holocene and Pleistocene; in older layers, judging from available data, it is less. It weakens quite considerably in all central parts of the region. Apparently, the main reason for this equalization, which, naturally, is not complete, is connected with modifications that organic matter undergoes when it is transferred from the place where it is produced to central and deep layers and after burial in the sediment in the course of early diagenesis. It is known that this is a general phenomenon, and investigations of sediments have given additional confirmations for practically all areas of the northern part of the Pacific Ocean. This question was specially considered by K. Emery [37] and O. K. Bordovskii [6, 7]. The distribution of organic matter in contemporary (Holocene) sediments of the Pacific Ocean is fairly complex. This is indicated by materials covering the sedimentation of organic carbon in most of the oceanic areas studied [8, 9, 28, 30, etc.]. The northern part of the Pacific Ocean, perhaps, reflects this important feature of oceanic sedimentation to the greatest degree. It is traced along the whole periphery of the region, which is characterized by very heterogeneous conditions, and to a lesser degree in its central part, where abyssal basins and ridges are located. At times, the bottom geomorphology is one of the most important factors in the succession of facies at great depth [27] and formation of sedimentation situations, on which the accumulation of organic matter on the bottom depends.

PECULIARITIES OF THE MANIFESTATION OF LATITUDINAL AND CIRCUMCONTINENTAL ZONALITY IN SEDIMENTATION OF ORGANIC MATTER IN THE NORTHERN PART OF THE PACIFIC OCEAN IN THE HOLOCENE

Although a general feature in the distribution of organic matter, as is known, is the relationship of this distribution with latitudinal and circumcontinental zonalities, in the northern part of the Pacific Ocean there are many regional differences. In essence, in each of the natural regions sedimentation of organic matter is determined by a direct dependence on the region's position in the system of latitudinal and circumcontinental zones and subzones. The action of individual factors is traced fairly sharply, for example, bottom relief, depth levels and transitions between them, and the amount and composition of sedimentary material. An important role belongs to vertical and horizontal circulation of waters, including up-

welling processes and water exchange between regions. Overall and local current systems show up everywhere, as do tidal phenomena and ices in certain regions. It is especially necessary to single out the biological productivity of waters, the vital activity of plankton and benthos as the main source of organic matter, the paths and scale of arrival of organic remains and their burial on the bottom, and the geochemical situation in sediments. Naturally all of these well known factors correspond differently to the types of zonality. Thus, in the northern Pacific-Ocean belt a transition from subpolar to boreal and subtropical regions is observed. Within these boundaries, a special place is occupied by the vast marginal seas of Northeast Asia, which are separated from the oceanic aquatory proper by island arcs. Water exchange between the seas and the ocean takes place through shallow (for example, the Sea of Japan, the eastern part of the Bering Sea) and deep-water (other seas) straits. In the region of Southern Alaska, special conditions are created by the central-Alaskan oceanic circulation, continental glaciation, and delivery of primarily fine material directly to the ocean by glaciers and rivers. Further south, in the Oregon and California zones, large-scale upwelling operates actively, with a complex system of cold and warm currents. As regards the greater part of the oceanic aquatory east and south of the island arcs from Japan to the Aleutians, it is occupied by waters of a high-latitude cyclonic circulation and the powerful currents bounding it: the warm Kuroshio and Northern Pacific-Ocean, and the cold Oyashio current.

Speaking of the role of circumcontinental zonality, one cannot fail to note that, as in other regions, as applied to sedimentation of organic matter it must be considered on different scales (macro, meso, and micro). Its specific features are also determined primarily by regular, though very complex changes in the sedimentary cover with distance from the shore and increasing depths. Superposition of macro-, meso-, and microscales takes place in the distribution of sediments by their lithological characteristics and genesis. As a result, large changes in the sedimentary process and, accordingly, in accumulation of organic matter are observed everywhere in the whole northern part of the Pacific Ocean. Thus, if the minimum amounts of organic carbon in sediments of the region are tenths and even hundredths of a percent, then the maximum amounts reach 5% or more. Two zones of minimum concentrations are traced. The first one encompasses coastal shallow waters, where coarse-grained, mostly terrigenous deposits are concentrated, while in regions adjacent to island arcs these are most often volcanic sands. As a rule, they occupy comparatively narrow strips of the bottom with increased steepness and dissection. The second zone of least concentrations looks completely different. It is vast in area and occupies large depths; the sediments are clayey silts, close to red clays. Two zones can also be noted in the distribution of the greatest concentrations of organic carbon in the northern part of the Pacific Ocean. The first is the zone of development of diatomic sediments in marginal seas and gulfs, and sediments in upwelling areas. It is not unified. The second is the zone of oceanic diatomic sediments, which is most significant in the northwestern part of the region, less so in the north and northeast. It is split up into parts and not continuous everywhere.

Circumcontinental zonality, as in other regions of the ocean, is viewed most fully on a macroscale. This is easily seen in the position of isolines of organic carbon in the region as a whole and in marginal seas. Attention should be paid to the fact that outside the underwater margins of the continents in the direction of greater depths the isolines of low content of organic carbon are usually much further apart than in the direction of the coast. The reasons for this are known; they are determined not only by geomorphological peculiarities of the corresponding marginal regions, and the distribution and composition of abyssal sediments, but also, what is particularly important, by a dependence on the system of bioproduction processes in the open ocean, arrival of organic remains at the bottom, and their burial.

Circumcontinental zonality on a mesoscale is traced everywhere. One of the clearest manifestations for organic carbon is its accumulation in belts of diatomic deposits in marginal seas (Bering, Okhotsk, Japan), where sedimentation of organic matter is especially significant. The same can be said of adjacent depths of the northwestern part of the Pacific Ocean. It is most fully expressed in the greater part of the deep-water basin of the Sea of Okhotsk [3]. In the basins of other marginal seas, accumulation of organic carbon in diatomic sediments is also not the same and, as a rule, diminishes with an increase in depths and intensification of the bottom oxidation layer in the water mass. The basins of the Sea of Japan and the Bering Sea serve as the clearest example [11, 12, 19, 20, 29]. This has also been established for the Kurile-Kamchatka region [1, 4]. Oceanological and biological prerequisites for such a change in concentrations of C_{org} have been discussed by many researchers [5-7, 29, etc.]. Mesoscale circumcontinental zones in the distribution of organic carbon are seen in deposits of the Gulf of Alaska [14], the East China Sea [15], and the California region [37]. Meso- and microscale manifestations of circumcontinental zonality are most fully indicated by summary data on the distribution of organic carbon in surface sediments of the oceanic depths as applied to all of the main regions of the Pacific-Ocean north, which are examined by E. A. Romankevich [29]. This also follows from the regional comparison of contents of C_{org} in sediments

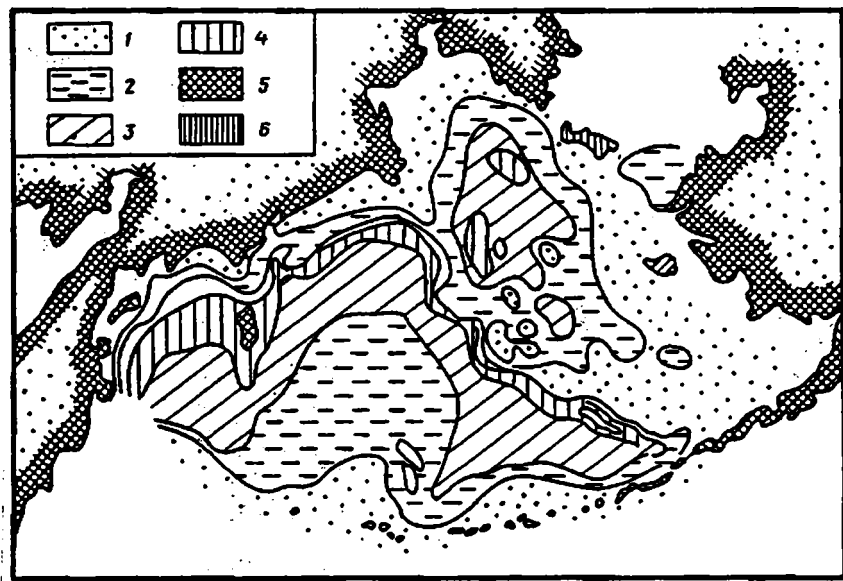


Fig. 4. Distribution of C_{org} in Holocene sediments of the Bering Sea, %: 1) <0.5 ; 2) $0.5-1.0$; 3) $1.0-1.5$; 4) $1.5-2.0$; 5) $2.0-2.5$; 6) >2.5 .

in a generalized form conducted by V. Ya. Trotsyuk and M. M. Marina [33]. But, if quite variable manifestations of circumcontinental zonation are observed in marginal seas and near the underwater margins of continents against an increased background of their concentrations ($0.5-2.5\%$), then in the depths of the ocean the pattern is different. The general background concentration of C_{org} there is much lower ($<0.5\%$, and frequently less than 0.25%). The areas covered by individual zones and subzones with low contents of organic carbon in sediments are usually large, and the transitions between them are always gradual.

Naturally, in this case it is impossible not to take into account the action of latitudinal and circumcontinental zonation, which occurs in all regions and determines at times a fairly complex pattern and at times even a variegated, though completely regular pattern of zonal distribution of organic carbon in contemporary (Holocene) and Pleistocene sediments (Figs. 4 and 5).

Within the framework of manifestation of natural zonation in individual regions, it is interesting to single out the factors that determine, to the greatest degree, the main features of accumulation of organic carbon in sediments. Thus, in the Bering Sea one has to note the special role of geomorphological and circulation factors. Among the former, we can emphasize the predominance of large depths in the western part and in the center of the sea, the widespread development of the shelf zone, especially south of the Bering strait and in the eastern part of the sea, and the subdivision of the steep continental coast by canyons. The sea is divided by extensive underwater ridges into deep-water basins, which are similar to oceanic ones to a known degree. This alone already creates conditions for inhomogeneity of the sedimentary process and the distribution of sediments. The role of the Aleutian island arc and straits freely connecting the Bering Sea with the Pacific Ocean and largely determining the oceanic effect on many sedimentation processes occurring in the Bering Sea and its organic world shows up to no small degree. In the Bering Sea, the boundary between the polar and boreal zones is sharply expressed; it runs not only in the northern part, but also in the east and is of great significance for dividing the waters and biocenoses, and also for the sea's bioproductivity. Intensive circulation of waters, especially along the continental slopes and in the regions of straits, promotes increased production of phytoplankton, especially diatomic plankton, which is the main source of organic matter for bottom deposits. It has recently been established that the high biological productivity of plankton and benthos of the eastern Bering shelf also affects the intensified supply of organic carbon to bottom sediments at small and large depths [12, 20, 39]. Large canyons in the slope promote removal of organic particles. The maximum accumulation of organic carbon has been established in diatomic sediments in fiord bays in the northwestern part of the sea (C_{org} up to 7%).

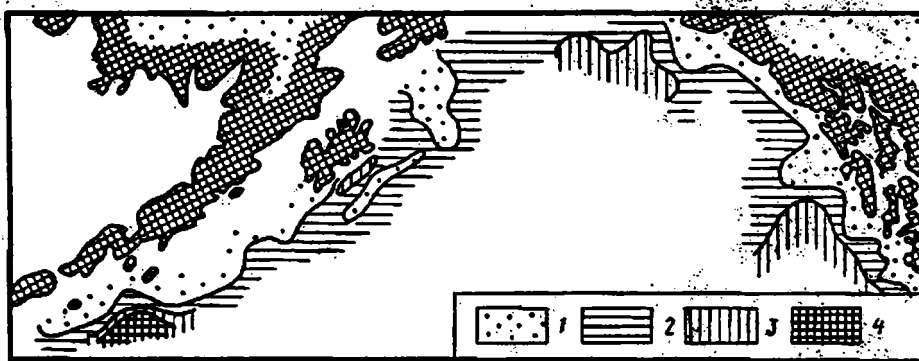


Fig. 5. Distribution of organic carbon in Holocene sediments on the shelf and slope of the Gulf of Alaska, %: 1) <0.5; 2) 0.5-1.0; 3) 1.0-1.5; 4) >1.5.

In the Sea of Okhotsk, many features of similarity with the Bering Sea are traced in the conditions of sedimentation of organic carbon. The Kurile island arc complicates the sea's bottom relief most of all. At the same time, the sea is distinguished by less intensity of water exchange with the Pacific Ocean. As a consequence, the effect of climatic and hydrological conditions grows, and the action of the continental vicinity, ice, and tidal phenomena shows up more strongly. Mixing of waters and their vertical circulation occur more actively. In a significant part of the sea's aquatory, diatomic plankton is produced intensively, the remains of which are deposited in a large amount in sediments of the deep-water basin north of the Kurile Islands and in a depression of the shelf in the northern part of the sea, where this is promoted by the circulation of currents and where there is no strong removal of particles settled on the bottom. The overall scheme of currents in the sea favors the entry and preservation of fine planktonogenic material in its center, while vertical circulation of waters promotes its constant reproduction. It is precisely on account of dying diatomic plankton that the sea's bottom sediments are so noticeably enriched with organic carbon. The relatively small runoff from dry land shows up in the entry of organic material of continental origin into bottom sediments. Even in bays in the western and northern parts of the sea it is not so great, and tides promote its removal into open regions of the sea, in which deposition of diatomic particles is sharply predominant.

Complex bottom relief is very characteristic of the Sea of Japan. The connection with the ocean is through shallow and narrow straits. Vertical circulation of waters encompassing the sea's whole mass is very significant. As a result, enrichment of the waters with oxygen, even in bottom layers, far exceeds that which takes place in the Sea of Okhotsk and Bering Sea with deep-water straits. The conditions of bioproduction differ sharply in the northern shallows of the Sea of Japan and in its deep-water basins. The interaction of warm currents, which are branches of the Kuroshio system's oceanic waters, with colder waters penetrating into the center of the sea from the north, and also as a result of the rise of cold waters with a decrease in depth on the basins' slopes, creates fairly sharp frontal divisions, where the neogenesis of diatomic plankton increases, as does its deposition on the bottom and, accordingly, certain enrichment of deep-water sediments with organic carbon. At the same time, the sea's oxygen and temperature conditions are such that with an increase in depths they promote decomposition of organic remains of plankton; accumulation of them there is not so abundant as in the Sea of Okhotsk and Bering Sea, although vertical circulation of waters apparently occurs more intensively and diversely. Hence the lesser degree of accumulation of organic carbon in waters of the Sea of Japan than in the Sea of Okhotsk and Bering Sea. In shallow waters, the significance of benthic biofilterers in enrichment of sediments with organic carbon rises, which frequently determines the pattern of its distribution on the shelf, and sometimes even on the continental slope.

In the shallow Yellow Sea, the main suppliers of organic matter are rivers, the flow of which drains the loess plains of China. This also takes place further south, in the vast shelf part of the East China Sea and more northern regions of the South China Sea. The pattern is different in areas with sharply increased accumulation of organic matter in contemporary sediments on account of high productivity of marine plankton and benthos. The main accumulation of C_{org} occurs as a result of its introduction with fine sedimentary material from dry land. This was a subject of attention in the works of F. Shepard et al. back in 1949 and in the later publications of H. Niino and K. Emery [38]. As a rule, the predominant concentrations of C_{org} rarely exceed 1-1.5%. And only in sediments on the continental slope of deep-water basins, where

vertical circulation, the rise of deep waters, and, accordingly, production of phytoplankton are intensified, does the amount of C_{org} increase to 2% or more [15].

In the South China Sea, organic matter accumulates more complexly, which is a result of the heavily dissected relief, unevenness in the entry of river sediments, intensive carbonate formation, coral reefs along the coasts and in the center of the sea, with their increased biological productivity, and abundant coastal fauna and flora. The subtropical conditions there, as well as somewhat further north, do not lead to a noticeable intensification of sedimentation of organic matter, rather the contrary.

Thus, in the marginal seas of Northeast Asia, practically everywhere in the sedimentogenesis of organic matter we see superposition of zonal changes from north to south, mainly connected with latitudinal geographic zonality, and from west to east, determined by the complex circumcontinental nature manifested in relief, climate, hydrology, and marine flora and fauna. The systems of island arcs and marine deep-water basins give sedimentation processes, including sedimentation of organic matter, especially great heterogeneity in these conditions. Perhaps it reaches a maximum in the region of the Sea of Japan.

With the transition to oceanic regions, such heterogeneity weakens. This is graphically indicated by O. K. Bordovskii's [6, 7], E. A. Romankevich's [29], and V. Ya. Trotsyuk and M. M. Marina's [33] summary data. Low contents of organic carbon in the sediments of island regions are fairly quickly replaced by increased concentrations toward the open ocean [3, 4] within the bounds of the Japanese Islands and, to a lesser degree, south of the Komandor and Aleutian Islands. There the content of C_{org} in deep-water sediments sharply exceeds 1-1.5% in most cases. Outside the belt of diatomic deposits, further toward central regions of the ocean, it drops to 0.25-0.5% and sometimes even smaller values, as far as one can judge from the available data, which is far from complete.

It is extremely significant that variations in the concentrations of C_{org} , when they take place at great depths, largely depend on the degree of dissection of the relief and variability of the sedimentary cover, even if these concentrations are minimal. However, the overall background of contemporary oceanic accumulation of C_{org} north of 30° north latitude is more or less consistent, which is reflected in composite maps of the distribution of C_{org} in open regions of the Pacific Ocean [29, etc.]. A certain increase is traced on the slopes of deep-water uplifts, the peaks and slopes of underwater mountains, and in deep-water channels.

Approaching the continental margins in the northeastern part of the Pacific Ocean, the amount of organic matter rises, and in sediments at the foot of the continents it exceeds 1-1.5%. There is a sharp increase in accumulation of organic carbon in the coastal Oregon-California zone, where the upwelling process becomes the leading factor in production of organic material. The strip of this zone is several hundreds of kilometers wide and extends more than 4000 km. The content of C_{org} varies from 1 to 1.5%, and even more in areas with the maximum manifestation of upwelling. K. Emery gives data for regions that he classifies as the so-called continental borderland, where the concentrations of C_{org} significantly exceed 2-3% [37]. In this region, development of the deep-water deposits most enriched with organic matter is observed. In the Gulf of Alaska, the field of such deposits encompasses its eastern and northeastern parts; however, it is not expressed along the coasts. There and further north and northwest, where the Alaskan current turns into the Aleutian current, the accumulation of C_{org} decreases. The whole sedimentation process is under the influence of polar conditions in the south of Alaska, large continental glaciers, and intensive removal of fine glacial-erosion material of the "glacial milk" type into the ocean.

Sedimentation situations with high and low accumulation of organic matter are quite close together; the contrast in their distribution is a distinctive feature of this region, which is largely promoted by the dissected bottom relief and interaction of warm and cold currents with different levels of biological productivity [12].

SEDIMENTATION OF ORGANIC MATTER IN SEDIMENTS IN THE NORTHERN PART OF THE PACIFIC OCEAN IN THE PLEISTOCENE

As shown by data on the content of organic carbon in sediments of the Pleistocene obtained in studying ground columns and cores from deep-water drilling, many common patterns in the sedimentation of organic matter in the northern part of the Pacific Ocean established for the Holocene are also traced in the Pleistocene. The role of latitudinal and circumcontinental zonality was manifested on the periphery, as well as in the center of the region. At the same time, the

TABLE 1

Borehole number	Sea depth, m	Borehole depth, m	Pleistocene thickness, m	Content of C _{org}
<i>Bering Sea</i>				
190	3875	627	152.9	0.3-0.7 (max. 0.7, hor.* 15, 41.2, 77.2)
186	4522	928	292.3	0.5-0.9 (max. 0.9, hor. 6, 7-15,
185	2110	728	65.1	167-169; min. 1165, 202.3)
188	2649	638	286.7	0.5-0.3 (max. 0.8, hor. 1)
<i>Pacific Ocean</i>				
192	3014, east of Kamchatka	942	56	0.3-0.4
183	4708, south of the Aleutian Islands	516	80.1	0.2-0.5 (max. 0.5, hor. 4.2, 6.2, 34.2, 51.1, 60.1; min. 0.1, hor. 60.9, 71.9)
179	3781, Gulf of Alaska	109	62.1	0.2-0.5 (max. 0.5, hor. 3.1, 25.6; min. 0.2, hor. 55.2, 62.1)
178	4218, Gulf of Alaska	794.5	197.4	0.3-0.6 (max 0.6, hor., 28.6, 34.6; min. 0.3, hor. 153.1, 175.4)
36	3273, northern California	116	44.2	0.3-0.4 (max. 0.4, hor. 16.1; min. 0.3, hor. 26.7, 32.2, 44.2)
47	2689, Shatskoe uplift	129	22.9	0.2-0.4

*Core horizons in meters.

succession of sediments characteristic of the Pleistocene, with different degrees of accumulation of diatomic material and, evidently, different rates of arrival and burial of organic matter [16, 29], is not expressed in the same way. In this regard, characteristics of the contents of C_{org} in Pleistocene deposits in the Bering Sea [6, 7, 12, 19, 20], Sea of Okhotsk, Kurile-Kamchatka region [1, 3, 4], Gulf of Alaska [14], and other adjacent regions in the northern part of the Pacific Ocean [29, 42] are quite indicative.

We can give as an example certain recent data on materials from deep-water drilling (Table 1).

In many boreholes located in the northwestern and northeastern basins with predominant depths of more than 4000 m, the thickness of the Pleistocene usually does not exceed several tens of meters, and the concentration of C_{org} is low (0.1-0.2%). Vertical variations are extremely small [33].

Data obtained from drilling materials in the Sea of Japan (voyage 127) are quite indicative. The variability in composition of Quaternary deposits there is perhaps the greatest for the northern part of the Pacific Ocean, and this is well traced in the content of C_{org}. Thus, for borehole 795 (bottom depth 3311.2 m) the total thickness of the Pleistocene is somewhat less than 80 m. The maximum content of C_{org} is 4.31% (depth 54.1 m); the minimum is 0.15% (depth 21.29 m). Sharp increases in content of C_{org} were established in horizons 4.61, 6.12, 22.23, 30.63, 37.23, and 47.16 m, where it exceeds 3%, which is much higher than in Holocene sediments. Low (0.2-0.3%) concentrations of C_{org} are also fairly frequent. The alternation of increased and low concentrations of C_{org} creates a special background of Pleistocene accumulation of organic matter in the Sea of Japan (Fig. 6a). It should be noted that, to a known degree, such a pattern is also observed for sedimentation of organic carbon in pre-Pleistocene time, which is undoubtedly significant for understanding certain peculiarities of the sedimentary process and paleogeography not only directly in the region of the Sea of Japan, but also, apparently, in adjacent regions. It may be that some of the reasons for such variations in sedimentation of organic matter in the Sea of Japan are the special nature of water exchange between the Sea of Japan and the ocean, repeated changes in its fluctuation, and their action on biological productivity. This does not take place in other regions, although many data also show variability in Pleistocene sedimentation of organic matter. O. K. Bordovskii [6], for example, gives the following results of his investigations in a deep-water area in the western part of the Bering Sea (Table 2).

TABLE 2

Station 540		Station 619	
Horizon, cm	C _{org} , %	Horizon, cm	C _{org} , %
4—10	1,2	20—40	1,57
60—70	0,93	125—145	1,05
545—560	0,87	215—230	1,12
740—755	0,78	280—304	1,00
923—943	0,91	440—460	0,61
1620—1645	0,83	670—690	0,71
		865—885	0,68
		1250—1260	0,61
		1340—1350	0,71

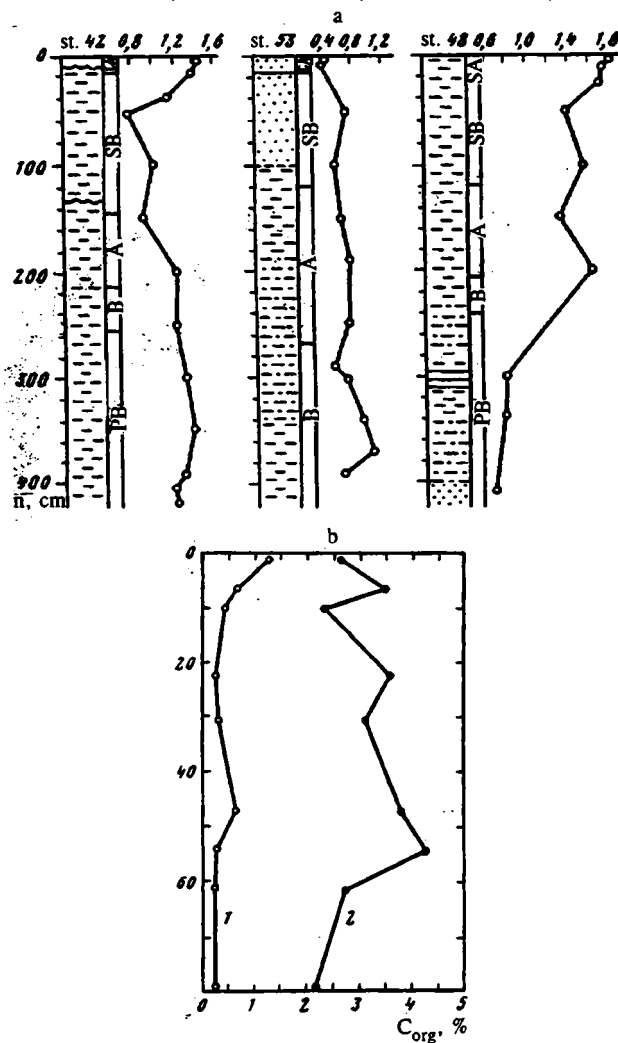


Fig. 6. Vertical distribution of C_{org} in Pleistocene sediments, %: a) in deep-water basin of the Sea of Japan ([42]); b) in the Navarin region of the Bering Sea [33]: 1, 2) horizons with minimum and maximum contents of C_{org}, respectively. Age indices: SA — sub-Atlantic, SB — sub-Boreal, A — Atlantic, B — Boreal, PB — pre-Boreal. Particle size of sediments - traditional notations.

It is noteworthy that at station 540, according to the results of diatomic analysis, no fewer than four horizons are established in Pleistocene layers [16], although with respect to the vertical distribution of C_{org} we can talk about three with a certain conditionality. The pattern is approximately the same at station 619. In the Navarin region of the Bering Sea, V. Ya. Trotsyuk and M. M. Marina note a quite similar pattern (Fig. 6b).

TABLE 3

Station 1019 (depth 3920 m)		Station 1081 (depth 3940 m)	
Horizon, cm	C _{org} , %	Horizon, cm	C _{org} , %
0—5	0,48	0—4	0,52
15	0,18	29	0,28
59	0,35	63	0,57
200	0,23	162	0,32
265	0,36	189	0,50
360	0,27	301	0,46

As E. A. Romankevich points out, in oceanic regions at depths exceeding 3800–4000 m, in layers that can be classified as Pleistocene the variability in concentrations of C_{org} is expressed differently in marginal and central parts of deep-water basins. For the former, the scale of variations in C_{org} is from 0.3 to 0.6%, sometimes somewhat more. Data for some deep-water drilling on the continental shelf west of Northern California are quite graphic. In borehole 32, at a depth of 4758 m, the thickness of the Pleistocene is small (7.9 m). The overall background of C_{org} content is 0.3–0.4%. It is higher at the surface (0.8%). In another, more northern borehole 33, at a depth of 4284 m, the Pleistocene has a thickness of 54.99 m, and the concentration of C_{org} is 0.4–0.5%, i.e., it practically does not change; in just one of the horizons at the boundary with the Pliocene does it rise to 0.6%. A similar background can be judged from our data obtained in the Alaskan abyssal plain [13] (Table 3). For sediments of the deep-water red-clay type developed at depths of about 4900–5000 m or more, the content of C_{org} is in the range of 0.2–0.4%, and variations, as a rule, are very small. On the whole, the concentration of C_{org} is quite consistent and indicates a noticeable equalization in the entry and transformation of organic matter at great depths in the Northern Pacific-Ocean region. But data on early diagenesis and its influence on deep-water sedimentation of organic matter still indicate that it takes place differently and depends on many factors determining the lithology of sediments, the composition and rate of arrival of organic remains to the bottom, regional peculiarities of bioproductivity, and the biogeochemical situation in individual regions. If the general features of this process were analyzed in N. M. Strakhov's well known publications, then we find specific consideration of them as applied to the northern part of the Pacific Ocean in works on the sedimentation and geochemistry of organic matter by O. K. Bordovskii, A. P. Lisitsyn, E. A. Romankevich, K. Emery, and other researchers.

THE ROLE OF VERTICAL CIRCULATION OF WATERS IN SEDIMENTOGENESIS OF ORGANIC CARBON IN QUATERNARY SEDIMENTS IN THE NORTHERN PART OF THE PACIFIC OCEAN

Investigation of organic carbon in Quaternary deposits in the northern part of the Pacific Ocean prompted researchers to pay special attention to the noticeable role of vertical circulation of waters in many features of sedimentation of C_{org} in the region. These features are manifested to the greatest degree on a mesoscale.

Naturally, in the light of its role in accumulation and distribution of C_{org} in sediments, vertical circulation of waters cannot be considered apart from other sediment-forming factors. But, at the same time, it is directly and indirectly reflected in the multifactor nature of sedimentation and promotes intensification of the unevenness and contrast of various situations of sedimentation of organic carbon through variability in the composition of flora and fauna, the overall productivity of organic matter, and its delivery to the bottom as a result of direct precipitation and biofiltration. One cannot fail to pay attention to the fact that vertical circulation of waters, in connection with changes in climate and hydrological conditions, which were at times fairly sharp in the northern part of the Pacific Ocean, is differently reflected in the production of organic matter and its subsequent migration. Even with known similarity of conditions in specific climatic and hydrological zones, it is possible to trace differences on the periphery and in central regions, and different degrees of consistency in the production and accumulation of organic material and its burial in sediments.

Looking at the position of zones of increased accumulation of C_{org}, one cannot fail to note that a significant part of them is located not only in regions of depressions in the relief and circular currents, where fine auto- and allochthonic sedimentary material and, accordingly, organic matter are always concentrated, but also in regions of intensive vertical circulation, particularly where subsurface waters rise up. It is known that these zones are due to frontal interaction of

currents, i.e., they are based on the hydrological nature and climatic prerequisites, or are the result of topographic action on the dynamics of waters and the whole course of the bioproduction process, deposition of organic matter on the bottom, and its burial in sediments. From the point of view of delivery of organic carbon to the bottom in sedimentary material, the role of topography and the associated rise of waters, especially in conditions of dissected relief, deserves no less attention than conventional ideas about the relations between hydrodynamics and accumulation of C_{org} in fine sediments. The topographic effect of a rise of waters, the consequence of which is enrichment of surface layers with biogenic elements and plankton and, when it dies, enrichment of bottom sediments with C_{org} , is fairly widespread in the northern part of the Pacific Ocean. Of course, it is not as significant in its scale of manifestation as the role of upwelling in the Oregon-California region, but, if we take the consequences of vertical circulation as a whole on the sedimentogenesis of organic carbon, then with all of their local differences and changes they are almost everywhere reflected in the actual pattern of entry and distribution of organic carbon in Quaternary sediments. Judging from the latest data, neogenic organic matter settles mostly where it is produced. There is no long-distance transfer of it in most regions. There is a clear interconnection (spatial and temporal) between zones of intensive vertical circulation of waters and zones with the greatest entry of organic matter into sediments and accumulation of increased concentrations of C_{org} in them. This process has been analyzed in detail on many examples in upwelling regions [36]. Researchers emphasize that vertical circulation of waters, as a mechanism of active transportation of organic matter, is important not only for raising the productivity of the photic layer, but also for speeding delivery of the remains of organisms after their death to the surface horizon of bottom sediments. This is connected with the fact that fine particles of organic matter are absorbed by plankton in the process of biofiltration of waters and increase their hydraulic size as a consequence of entry into the composition of pellets or detrital sediments and settle on the bottom at a much higher rate than with simple precipitation in the form of a fine suspension together with clay material. Moreover, the process of downwelling also participates in the vertical transfer of organic matter: descent of waters that occurs in canyons on the continental slope and also accelerates the delivery of organic particles within the bounds of continental and island slopes [24]. Thus, vertical circulation of waters, which is determined to no small degree by dissection of the relief, is a significant element of the system increasing production of organic matter and the rate of its entry into sediment.

Apparently, we can talk about several types of vertical circulation of waters affecting the production and transfer of organic particles in the region. Naturally, the most important and significant is the upwelling type. Next in significance comes the rise of waters along long continental slopes. It is well expressed in the western and eastern parts of the Bering Sea, and along Kamchatka. In the northeastern part of the Northern Pacific-Ocean region a sort of superposition of this type of vertical circulation on upwelling takes place. The nature of it is fairly complex and cannot be reduced just to the peculiarities of bottom topography.

The overall circulation of waters in zones of the continental slope, local differences in the distribution of plankton and benthos, and frontal interactions, which are sometimes associated with the slopes' general topography, are evidently of major significance. It should be noted that data obtained in this regard as a result of investigation of bottom deposits on the eastern Bering-Sea slope help, in particular, in evaluating more thoroughly the role of continental slopes and the vertical circulation of waters associated with them in sedimentogenesis of organic matter in the sea [12, 14]. In the Sea of Okhotsk, and to a known degree at the boundary between the northern and central parts of the Sea of Japan, due to peculiarities of the mesorelief we encounter a somewhat modified type of vertical circulation of waters, which is manifested in action on the productivity and accumulation of C_{org} in sediments not in narrow zones but, on the contrary, over a relatively wide aquatory, where regions of development of diatomic sediments of considerable area are formed. To a known degree, a similar pattern can also be traced in deep-water basins of the Bering Sea, where bottom deposits with high contents of C_{org} lie in the lower, flattened part of the continental slopes and at their foot. The nature of vertical circulation also acquires certain specific features in so-called nodal regions, where sections of the continental slope with different trends come together, and the whole dynamics of waters and structure of the sedimentary process is complicated, including production and sedimentation of organic matter. The southeastern part of the Bering Sea, where the Umnak plateau is located, the northwestern part of the sea south of the Anadyr Gulf, sections of the slope's conjunction with the Shirshov ridge, and some others can serve as examples.

In the zones of island arcs, the manifestation of vertical circulation of waters differs somewhat as applied to productivity and accumulation of organic matter. Sometimes, its action is not intensified, but even weakens. This may depend on a number of factors, including the degree of mixing of waters in interisland straits, the direction and speed of currents, and forms of mesorelief, including sills. On the marine slopes of island arcs, the pattern is frequently different.

There, as on individual slopes of deep-water trenches with stepped relief (Aleutian, Kurile-Kamchatka, Ryukyu), vertical circulation of waters promotes increased accumulation of organic material [2, 15, 34].

One more type of vertical circulation of waters promoting an increase in production processes and accumulation of C_{org} in sediments is the local rise of waters on deep-water elevations (the Shatskoe uplift), underwater mountains (the Gulf of Alaska, Emperor mountains, Hawaiian ridge, South China Sea), and their slopes. This results in a certain variegation in the distribution of C_{org} at great depths in abyssal basins.

Thus, vertical circulation of waters associated with the relief is quite significant and complicates the sedimentation of organic matter in deep-water sediments in the northern part of the Pacific Ocean. Naturally, it is manifested against the overall background of large-scale climatic and hydrological processes occurring in the Holocene and Pleistocene, "modeling" them. Significant changes in the overall course of Quaternary sedimentation caused by the succession of glacial and interglacial epochs, which are especially well established at the boundary between the Holocene and Pleistocene, did not lead to fundamental changes in the course of vertical circulation of waters and the factors causing them with the transition from peripheral to central regions. In contrast to a number of other characteristics of bottom sediments, accumulation of C_{org} preserved its overall scheme. For the period of great natural fluctuations which distinguish Quaternary time, the Neogene and Paleogene, this is a remarkable feature of the sedimentation of organic matter [26, 35].

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CALCIUM RHODOCHROSITE IN SEDIMENTS OF THE GUATAMALAN BASIN

V. N. Sval'nov and V. N. Kuleshov

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This article discusses data on the composition and distribution of diagenetic manganese carbonate in columns of Quaternary sediments of the Guatamalan basin.

Authigenic carbonates with a complex composition are fairly widespread in Pleistocene–Holocene sediments of lakes, seas, and the region of continental-margin lithogenesis in the ocean [1, 7-10, 14, 15, 17, 18, 20, 22, 23, 25]. They are characterized by variable contents of Mn, Ca, Mg, and Fe. In particular, it is believed [9] that favorable conditions for formation of authigenic manganese carbonate (rhodochrosite) are created in the process of diagenesis in the presence of low contents of organic matter in the sediments. Such conditions are observed in the ocean's transition zone from coastal (strongly reduced sediments) to pelagic (oxidized sediments). Rhodochrosite, consequently, is a mineralogical criterion of oceanic lithogenesis near continents [13]. New confirmation of this conclusion was obtained in the present data from the Guatamalan basin.

MATERIAL AND METHODS OF INVESTIGATION

The forty-first voyage of the research ship *Dmitrii Mendeleev*, carried out comprehensive study of the regional variability of fields of ferromanganese concretions, the sediments surrounding them, and sedimentation processes [12]. The Guatamalan basin was chosen as one of the regions for this work. The geological testing ground there had the following coordinates of the central point: 6°30' north latitude and 92°00' west longitude. Sediments were collected at depths of 3490-3740 m. In all, 16 dredge samples and 11 sediment columns of up to 412 cm were obtained, and two monoliths were lifted with M. A. Voll's sealed tube. Besides that, patterns of hemipelagic sedimentation were revealed on a submeridional profile from 8°38.6' north latitude 93°34.0' west longitude to 12°16.8' north latitude 96°06.7' west longitude adjacent on the north to the area of detailed work in the Guatamalan basin. On the profile, six dredge samples and six sediment columns up to 382 cm long were lifted, and one monolith with the use of the sealed tube.

Distinctive features of the part of the Guatamalan basin that was examined are: intensive enrichment of the upper 10-25 cm of sediments with manganese hydroxides, the presence of large ferromanganese concretions on the surface and in the strata, widespread occurrence of products of acid volcanism, a transition along the northern periphery of the testing ground from variegated sediments to gray-colored ones, their relatively high silicon content, and noticeably varying carbonate content. The surface (homogeneous) layer of the cross sections is represented by dark-brown, clay-radiolarian oozes, frequently enriched with diatoms, sometimes slightly calcareous. In places, radiolarian-diatomic sediments are found. The thickness of the homogenous layer varies in the range of 2-11 cm. Below, a variegated bioturbation layer is disclosed in sediments oxidized to various degrees. The predominant shades of the sediments' color are brown, light-brown, and pale yellow; at a number of stations, deeper than 30-70 cm lie yellowish- and greenish-gray oozes with a gradual transition. The base of the Quaternary cross sections that were disclosed is clay-radiolarian sediments, often slightly calcareous, in places enriched with diatoms. Layers of foraminifer-coccolith ooze enriched with radiolarians, and radiolarian-diatomic deposits are also found.

In the investigated columns, two levels of accumulation of volcanoclastic material are traced, which are represented by tuffite siliceous-clay oozes, and layers and lenses of light-gray tephra of acid composition. At most stations, tephra lenses or layers 2-6 cm thick, and also clay radiolarian and radiolarian-diatomic tuffite oozes 3-19 cm thick intersecting them lie within the bounds of the 5-44 cm horizon. The second level of enrichment of clay-radiolarian oozes with volcanoclastic material is traced in the interval from 80 to 310 cm.

Institute of Oceanology, Russian Academy of Sciences, Moscow. Translated from *Litologiya i Poleznye Iskopaemye*, No. 3, pp. 21-35, May-June, 1994. Original article submitted December 7, 1992.

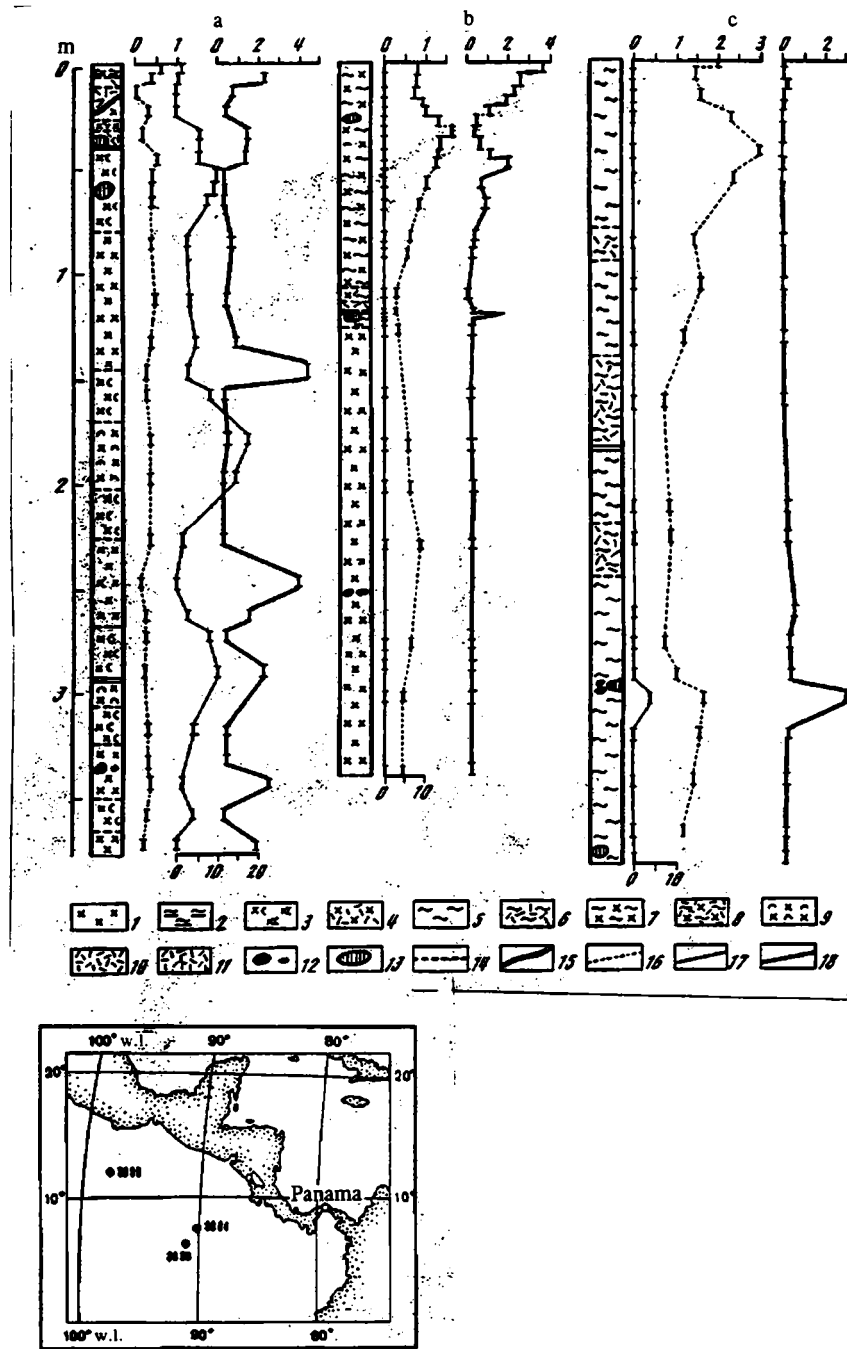


Fig. 1. Lithological and geochemical characteristics of sediments (a - station 3850, depth 3660 m; b - station 3884, depth 3585 m; c - station 3899, depth 4095 m). 1-9) oozes (1 - clay-radiolarian, 2 - the same, enriched with diatoms, 3 - the same, slightly calcareous, 4 - the same, tuffite, 5 - clay-diatomic, enriched with radiolarians, 6 - the same, tuffite, 7 - radiolarian-diatomic, 8 - the same, tuffite, 9 - calcareous-clay, enriched with radiolarians); 10) tephra; 11) the same, slightly calcareous; 12) nodules of calcium rhodochrosite; 13) ferromanganese concretions; 14, 15) interfaces (14 - gradual, 15 - sharp); 16-18) curves of contents, %: 16) C_{org} , 17) CO_2 , 18) Mn_{tot} .

On the submeridional profile, strata of greenish-gray and grayish-green sediments with various degrees of bioturbation were disclosed, which are colored brown (1-2 cm) and brownish-gray (down to 35 cm) near the bottom surface. The surface (homogeneous) layer of the cross sections is represented by clay-radiolarian oozes enriched with diatoms, and

radiolarian-diatomic sediments. The thickness of the homogeneous layer varies in the range of 5-11 cm. Columns were composed of diatomic, radiolarian-diatomic, diatomic-clay, and clay-diatomic oozes, frequently enriched with radiolarians. A fairly significant part of the disclosed strata is radiolarian-clay and clay-radiolarian sediments, sometimes enriched with diatoms or slightly calcareous. A characteristic feature of the studied columns is lenses and layers of tephra with an acid composition, and also tuffite siliceous-clay sediments intersecting them. The thickness of the ash layers varies from 3 to 10 cm; and of the tuffite sediments, in the range of 5-40 cm. The number of levels of enrichment of the oozes with volcanoclastic material reaches three-four.

Macro- and microscopic study of the specimens that were collected made it possible to reveal nodules and dispersed fine aggregates of authigenic manganese-carbonate crystals at stations 3850, 3884, and 3899 (Fig. 1). The nodules' chemical composition was determined on a Plasma-AFS atomic-fluorescent spectroanalyzer and by methods of classical chemistry (analysts G. A. Granovskaya and V. V. Karpushina). X-ray study of air-dry preparations was performed on a DRON-2 diffractometer with CuK_α radiation, graphite monochromator, Ni filter, 40 kV and 20 mA, and a slit of 0.5×0.5 mm (analysts Z. N. Gorbunova and E. V. Pokrovskaya). Complete silicate analysis of the sediments was done by T. G. Kuz'mina by the X-ray fluorescent method on a VRA-30 instrument. On an AAs-403 atomic-absorption spectrophotometer, Al, Fe, Mn, Ti, K, Ni, Co, Zn, Li, Cu, Pb, V, Cr, and Mo were determined in the sediments (analyst N. N. Zavadskaya). In analyzing CO_2 and C_{org} , N. P. Tolmacheva used the combustion method [11]. Electron microphotographs were obtained by V. A. Karlov on an ISM-U3 scanning microscope. Paleomagnetic investigations of the sediments were conducted by Yu. Yu. Ivanov. The cross sections were subdivided according to diatom complexes by V. V. Mukhina. To study the isotopic composition of carbon and oxygen, we used nodules of manganese carbonate, and also shells of plankton foraminifers extracted from nodules and the surrounding sediments.

The samples were prepared for mass-spectrometric analysis in a vacuum according to the standard procedure in orthophosphoric H_3PO_4 acid at room temperature. Samples of nodules and the surrounding sediments were decomposed in two stages. First, a sample preliminarily ground to fine powder was dropped into the acid and dissolved for 1 h. The CO_2 gas released during that time was collected and removed from the system. This portion of the gas was supposed to correspond to the decomposition product of calcite. Then the sample was left to finish decomposing in the system for 2 days. The gas released was also collected and removed from the system. The second portion of gas should correspond to the decomposition product of hard-to-dissolve carbonates [24].

Results of the mineralogical investigation of the studied samples show that, in our specimens, hard-to-dissolve carbonates are represented by manganese carbonate-calcium rhodochrosite.

It is obvious that the procedure used is not devoid of shortcomings. However, the authors believe that it is fully suitable for solving the problem posed, and the isotopic data obtained are sufficiently representative.

Mass-spectrometric measurements were made on a domestic MI-1201V mass spectrometer. Reproducibility of the measurement results in parallel tests was no worse than $\pm 0.3\%$.

RESULTS OF THE INVESTIGATION

Composition of Sediments. At station 3899 ($12^\circ 16.3'$ north latitude, $97^\circ 06.0'$ west longitude; depth 4095 m), the thickness of the disclosed Quaternary cross section is 382 cm. The column is mostly represented (see Fig. 1) by grayish-green and greenish-gray pelitic clay-diatomic oozes enriched with radiolarians. In the upper part of the cross section (0-1 cm) the color of the sediments is brown. In places, open passages of limivores are found. Below 120 cm, the oozes have noticeable bioturbation, and their coloring frequently becomes marbled, on account of dark-gray fringes of manganese hydroxides, enveloping round lightened areas - traces of bioturbation. At depths of 180-185 cm lie lenses of light-gray silty-psammitic tephra with acid composition. In the intervals of 80-95, 140-180, and 220-245 cm, clay-diatomic tuffite oozes are revealed (15-30% colorless volcanic glass).

Nodules of manganese carbonate up to 5 cm in size are found at depths of 295-300 cm. One of the nodules gradually changes to a decomposing ferromanganese concretion, from which practically all of the manganese has been removed as a result of diagenesis. The inner part of the nodules is dense and gray-colored; the outer jacket is loose and brownish-gray. Occasional skeletal remains of plankton foraminifers, radiolarians, and diatoms are found in the nodules. Directly above the horizon with the nodules (in the interval of 293-295 cm), fine aggregates of manganese carbonate amount to about 3% of the section's area; and below (300-302 cm horizon), as much as 45%.

TABLE 1. Chemical Composition of Basic Types of Studied Sediments (% of dry matter)*

Station number	Horizon, cm	Type of sediment**	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	MnO	MgO	CaO	K ₂ O	Calcination loss	Total
3850 (6°37.5' north latitude, 93°21.0' west longitude)	10—14	T	67,6	0,32	12,4	3,16	0,97	1,50	1,06	3,38	7,2	97,59
	30—35	Tsc	46,9	0,46	10,2	4,91	1,54	1,97	11,20	2,04	18,3	97,52
	130—135	Csc	47,5	0,52	9,3	6,08	1,02	3,08	8,22	1,36	20,5	97,58
	340—345	CR	50,0	0,56	10,1	6,19	2,80	3,53	5,18	1,47	17,8	97,63
3884 (7°27.0' north latitude, 92°43.8' west longitude)	0—5	RD	54,4	0,62	10,5	4,83	3,94	3,32	1,48	1,43	17,1	97,62
	12—16	"	48,3	0,56	9,7	5,27	2,27	3,22	6,25	1,35	20,6	97,52
	25—30	"	53,2	0,53	9,5	6,18	0,55	4,07	3,19	1,46	18,8	97,48
	55—60	"	54,7	0,49	9,5	5,03	0,86	4,14	2,74	1,67	18,4	97,62
	87—93	"	56,9	0,45	10,2	4,80	0,40	3,23	3,44	2,14	16,0	97,65
	119—120	CRd	55,5	0,62	11,0	6,20	2,04	3,53	1,37	1,63	15,7	97,59
	125—130	CR	59,1	0,57	10,6	8,08	0,42	2,96	1,38	1,85	12,5	97,46
	285—290	"	56,1	0,58	10,6	5,80	0,42	3,03	2,28	1,65	17,0	97,46
3899 (12°16.3' north latitude, 97°06.0' west longitude)	0—5	CDr	55,0	0,60	12,5	6,30	0,25	3,38	1,11	1,55	16,7	97,39
	25—30	"	54,0	0,74	14,0	6,11	0,13	3,25	1,81	1,94	15,5	97,48
	105—110	"	56,4	0,78	12,7	5,92	0,19	3,67	1,43	1,99	14,5	97,58
	260—265	"	56,2	0,66	13,1	6,34	0,61	3,61	1,25	1,80	14,0	97,57
	290—295	"	54,0	0,79	15,0	6,63	0,58	3,59	1,84	2,01	13,0	97,44
	340—345	"	56,3	0,91	14,3	6,00	0,23	2,83	1,40	2,10	13,4	97,47
	377—382	"	57,1	0,63	13,2	6,82	0,26	3,70	1,03	2,11	12,7	97,55

*Analyst T. G. Kuz'mina (Institute of Oceanography, Russian Academy of Sciences).

**T — tephra; Tsc — the same, slightly calcareous; CR — clay-radiolarian; CRd — the same, enriched with diatoms; RD — radiolarian-diatomic; CDr — clay-diatomic, enriched with radiolarians.

The column's clay-diatom oozes are characterized by fairly stable chemical composition (Tables 1 and 2). The content of residual organic carbon varies from 0.7 to 3.0% (see Fig. 1), diminishing downward through the cross section; however, relative subsurface maximums are observed at depths of 42-47 and 301-306 cm (3.0 and 1.6%, respectively). The sediments' carbonate content (figured in terms of CaCO_3) is less than 1%, and only directly below the horizon with the nodules does it reach 8.3%. There, the peak of manganese is revealed, against a background of consistently low concentrations, and a certain enrichment of the oozes with vanadium is traced. No anomalies are found in the behavior of the other analyzed elements. Their distribution in the cross section quite adequately reflects conditions of reducing diagenesis, and also changes in the types of sediments.

At station 3884 ($7^\circ 27.0'$ north latitude, $92^\circ 43.8'$ west longitude; depth 3585 m), a sediment column 340 cm long was obtained. The upper part of the cross section (0-114 cm) is represented by pelitic radiolarian-diatom oozes (see Fig. 1). In the interval of 0-11 cm, the sediments are dark-brown; lower (11-22 cm), yellowish spots are distinguished against this background. At a depth of 22-29 cm lie rusty-gray oozes that include a ferromanganese concretion 0.8×1.5 cm in size. Deeper, gray (to greenish-gray) sediments with admixtures and segregations of hydrotoilite are traced. At the base of the layer, dark-gray radiolarian-diatom tuffite ooze (105-110 cm horizon) is replaced by light-gray, also tuffite, pseudolayered sediment. In the interval of 114-125 cm, a thin-layered member is found: 114-115 cm — cemented gray tephra with psammitic-silty texture, bordered at the roof of the layer by a film of manganese hydroxides; 115-116 cm — yellowish-gray silty-psammitic tephra; 116-119 cm — grayish-rust-colored radiolarian-diatom tuffite ooze; 119-121 cm — brownish-gray clay-radiolarian sediment enriched with diatoms, which includes a ferromanganese concretion 1.5×2.0 cm in size; 121-125 cm — rusty-gray clay-radiolarian ooze.

The 125-340-cm horizon is composed of greenish-gray clay-radiolarian sediments, below 180 cm with a bluish tint, admixtures of hydrotoilite, and a slightly marshy smell. Dark-gray layers and spots are found at depths of 249-259, 265-266, and 285-290 cm. In the interval of 249-255 cm, among clay-calcareous ooze enriched with radiolarians, yellowish-gray, flattened nodules of manganese carbonate up to 1.5 cm in size are found. These contain as much as 15% radiolarian skeletons, 3% diatoms, and detritus of isolated plankton foraminifers. Fine aggregates of authigenic carbonate are also revealed at depths of 48-50, 60-62, and 119-121 cm (20, 2, and 3% of the section's area, respectively). The contents of plankton foraminifer detritus in the surrounding oozes does not exceed 1%.

In chemical composition, the sediments of station 3884 are somewhat impoverished with organic matter and aluminum in relation to the sediments of station 3899, but enriched with manganese (see Tables 1 and 2). The carbonate content of clay-radiolarian and radiolarian-diatom oozes is usually less than 1% CaCO_3 . The content of residual organic carbon varies in the range of 0.25-1.64%. The maximum is found at a depth of 30-35 cm. Comparatively increased concentrations of manganese are confined to sediments with segregations of authigenic carbonate (see Fig. 1), and its maximum (3.66%) is noted in the 0-5-cm layer. Clay-radiolarian oozes (119-121-cm horizon), which also contain a ferromanganese concretion, are enriched with cobalt and copper.

At station 3850 ($6^\circ 37.5'$ north latitude, $93^\circ 21.0'$ west longitude; depth 3660 m), the thickness of the disclosed Quaternary cross section is 376 cm (see Fig. 1). The column is mostly represented to varying degrees by oxidized pelitic clay-radiolarian oozes, sometimes slightly calcareous, tuffite, or enriched with diatoms. The color of the sediments varies from dark- to yellowish- and grayish-brown. As a result of bioturbation and redistribution of manganese hydroxides, a marbled pattern is quite often observed. At depths of 170-201 and 292-306 cm lie light-brown and pale-yellow calcareous-clay oozes enriched with radiolarians. The 8-22- and 30-36-cm horizons are composed of light-gray tephra with silty-pelitic and psammitic-silty textures. The lower ash layer includes a ferromanganese concretion 3×4 cm in size. One more concretion (7×7 cm) was found among slightly calcareous clay-radiolarian oozes at a depth of 56-63 cm. In the interval of 330-340 cm, light-brown clay-radiolarian oozes include yellowish-white nodules of manganese carbonate up to 3 cm in size on their long axis. Dispersed fine aggregates of authigenic carbonate are also found at a depth of 146-148 cm, where they amount to 3-5% of the section's area. The clay-radiolarian ooze surrounding them contains rounded segregations of manganese hydroxides (as much as 7%), detritus of plankton foraminifer skeletons (5-7%) and coccoliths (1-2%). In the nodules of manganese carbonate, isolated radiolarians and about 2% plankton foraminifer detritus are found.

The distribution of elements in the column (see Tables 1 and 2) is controlled by the types of sediments and diagenetic processes. The contents of residual organic carbon vary in the range of 0.20-0.66%, decreasing slightly down through the cross section. The sediments' carbonate content is extremely variable and rises in places from less than 1 to

TABLE 2. Chemical Composition of Sediments after Conversion to Carbonate-Free Dry Matter*

Horizon, cm	Type of sediment**	Al	Fe	Mn	Ti	K	Ni	Co	Cu	Zn	Li	Pb	V	Cr	Mo	CaCO ₃
Station 3850																
0—5	Crđ	5,71	4,16	2,88	0,30	1,60	552	49	405	334	31	10	102	13	<10	2,23
5—8	CRt	6,00	4,13	2,41	0,27	1,98	589	47	428	303	33	20	122	13	<10	1,50
10—15	T	6,39	1,99	0,66	0,20	2,37	210	32	380	106	26	43	36	22	—	<1
20—25	Cr	5,44	3,66	0,48	0,26	1,28	286	36	2410	194	19	52	73	52	—	<1
30—35	Tsc	5,24	4,31	1,70	—	—	482	30	168	245	—	—	—	—	—	14,14
42—48	CRsc	5,98	3,92	1,62	0,38	0,93	617	51	2409	310	17	44	11	47	—	14,09
52—56	"	5,72	4,96	0,38	0,47	1,00	345	34	1411	291	29	53	105	88	—	24,15
57—63	"	3,26	5,62	0,56	—	—	296	19	138	326	—	—	—	—	—	21,73
63—68	"	5,81	4,96	0,40	0,38	1,38	376	42	1685	305	41	56	125	78	—	18,11
83—87	CR	5,13	6,08	0,90	—	—	378	38	234	170	—	—	—	—	—	7,41
110—115	"	3,81	5,85	0,53	—	—	322	26	172	175	—	—	—	—	—	8,14
130—135	CRsc	3,18	5,45	1,16	—	—	383	27	218	221	—	—	—	—	—	1,98
143—150	CR	5,32	4,43	4,81	0,43	0,67	147	120	1237	525	19	19	166	37	—	7,04
155—160	CRsc	3,36	5,71	0,49	—	—	250	30	150	257	—	—	—	—	—	19,75
177—182	CCr	5,41	4,40	0,85	0,44	0,85	403	55	1205	221	26	79	74	75	—	40,25
195—200	"	4,97	5,22	0,62	—	—	304	22	113	427	—	—	—	—	—	32,59
224—230	CR	5,87	5,60	0,45	0,42	1,40	369	48	1813	231	34	46	115	65	—	4,03
245—250	"	5,21	4,94	4,08	0,44	0,72	130	114	1150	432	15	26	183	28	—	<1
260—265	"	3,67	5,42	1,77	—	—	467	38	259	195	—	—	—	—	—	7,41
270—275	CRsc	5,05	5,91	0,52	—	—	270	36	143	287	—	—	—	—	—	18,77
287—292	4	4,72	5,88	2,86	—	—	644	50	216	408	—	—	—	—	—	23,70
315—320	"	5,83	5,13	0,51	0,35	1,23	377	33	477	431	41	33	61	29	<10	9,89
330—335	CR	4,63	5,46	0,49	—	—	279	27	437	390	—	—	—	—	—	5,00
340—345	"	5,12	4,89	2,65	0,37	0,76	112	85	1021	389	20	44	139	153	—	5,03
355—360	CRsc	5,22	5,96	0,41	—	—	263	29	428	347	—	—	—	—	—	9,88
370—375	CR	5,14	5,00	1,86	0,40	0,64	762	52	1310	305	16	49	124	86	—	2,01

TABLE 2 (continued)

Horizon, cm	Type of sediment**	Al	Fe	Mn	Ti	K	Ni	Co	Cu	Zn	Li	Pb	V	Cr	Mo	CaCO ₃
Station 3884																
0—5	RD	4,86	4,02	3,66	0,27	1,16	750	44	445	609	33	<10	145	23	<10	<1
5—12	"	4,79	4,13	2,53	0,31	1,06	670	45	407	616	32	10	135	30	<10	<1
12—16	"	5,14	4,08	2,24	0,33	1,02	639	47	375	592	33	10	139	24	<10	3
16—20	"	5,16	4,21	1,91	0,30	1,11	560	47	366	594	35	10	135	33	10	2
20—25	"	4,70	4,47	1,17	0,30	1,04	440	35	355	587	32	10	95	23	10	1
25—30	"	4,53	4,32	0,51	0,25	1,22	490	32	272	670	37	10	120	26	10	1
30—35	"	4,88	4,14	0,42	0,30	1,43	680	39	499	703	41	10	205	30	<10	1
35—42	"	4,81	3,74	0,66	0,25	1,43	750	44	319	809	43	10	225	26	<10	1
42—46	"	5,05	3,91	1,25	0,27	1,39	840	49	435	721	44	10	150	26	<10	<1
46—50	"	4,68	3,42	2,01	0,24	1,26	800	46	259	686	41	10	130	17	<10	<1
55—60	"	5,08	3,61	0,81	0,26	1,43	830	50	336	642	40	10	85	23	<10	<1
65—70	"	5,25	3,50	0,95	0,25	1,61	730	44	409	589	41	10	95	33	<10	<1
80—85	"	5,52	3,47	0,36	0,23	1,87	370	30	216	401	42	10	60	36	<10	1
88—93	"	5,82	3,17	0,27	0,23	2,29	340	29	193	340	40	10	5	20	10	2
108—112	RDt	5,92	1,67	0,10	0,14	2,90	120	19	263	149	29	10	20	3	10	<1
118—119	"	5,45	4,40	0,38	0,24	1,61	310	28	308	302	35	10	60	13	<10	<1
119—120	CRd	5,75	4,56	1,69	0,27	1,26	720	62	1270	350	33	10	55	20	<10	<1
120—122	CR	5,58	5,60	0,38	0,27	1,56	350	32	336	333	40	10	30	49	<10	<1
125—130	"	5,52	5,22	0,30	0,27	1,87	400	51	590	381	43	10	70	53	<10	<1
155—160	"	6,52	4,87	0,18	0,36	1,65	340	32	293	346	51	10	105	66	<10	<1
180—185	"	4,68	3,74	0,18	0,23	1,09	280	28	418	464	56	10	275	66	<10	<1
200—205	"	5,45	4,73	0,28	0,27	1,35	420	29	261	515	53	10	190	36	<10	1
228—232	"	5,05	4,34	0,35	0,27	1,35	900	66	362	679	52	20	180	53	10	2
275—280	"	5,12	4,18	0,33	0,27	1,11	700	57	520	532	39	—	145	43	<10	<1
285—290	"	5,65	4,24	0,31	0,30	1,48	760	50	1450	508	46	30	145	59	<10	<1
300—305	"	6,22	4,95	0,27	0,33	1,61	400	29	370	416	54	30	85	56	10	<1
335—340	"	5,48	4,05	0,73	0,33	1,52	360	32	227	478	47	20	75	69	<10	1

TABLE 2 (continued)

Horizon, cm	Type of sediment**	Al	Fe	Mn	Ti	K	Ni	Co	Cu	Zn	Li	Pb	V	Cr	Mo	CaCO ₃
Station 3899																
0—3	CDr	6,92	4,62	0,16	0,27	1,52	170	18	206	379	53	20	115	46	<10	<1
3—10	"	6,52	4,40	0,10	0,25	1,61	200	20	255	609	53	20	140	40	10	<1
15—20	"	7,22	4,48	0,10	0,25	1,61	310	27	265	412	61	20	155	53	<10	<1
25—30	"	7,49	4,21	0,08	0,30	1,65	250	19	233	508	64	20	170	59	30	<1
42—47	"	6,89	4,26	0,08	0,28	1,48	250	25	184	375	58	10	150	49	<10	<1
55—60	"	7,15	4,29	0,07	0,29	1,75	200	24	163	353	63	10	155	53	20	<1
84—89	CDrt	6,72	3,50	0,09	0,21	2,11	160	24	139	394	54	10	105	36	20	<1
105—110	CDr	6,55	4,10	0,14	0,25	1,91	240	27	197	438	63	10	150	63	<10	<1
130—135	"	6,82	3,83	0,14	0,21	1,91	410	32	208	390	56	20	160	53	20	<1
160—165	CDrt	6,12	2,79	0,14	0,15	2,22	400	23	233	318	43	10	135	20	20	<1
210—215	CDr	6,79	4,40	0,21	0,28	1,78	220	19	274	464	64	<10	155	36	<10	<1
225—230	CDrt	6,45	4,10	0,18	0,25	2,16	260	22	293	440	58	10	180	40	20	<1
260—265	CDr	6,55	4,48	0,46	0,27	1,48	340	22	336	447	58	10	145	36	<10	<1
275—280	"	7,05	4,42	0,28	0,27	1,48	290	22	336	454	60	10	155	46	10	<1
290—295	"	7,72	4,40	0,43	0,28	1,61	180	20	160	381	69	10	145	59	20	<1
301—306	"	7,44	4,11	3,33	0,36	1,76	283	24	194	358	77	11	207	69	22	8,28
318—323	"	7,36	4,56	0,27	0,27	1,69	210	24	182	258	74	20	145	66	20	<1
340—345	"	7,19	4,13	0,16	0,30	1,73	140	18	167	267	77	10	180	63	10	<1
365—370	"	6,85	4,21	0,19	0,27	1,65	420	26	225	335	65	10	175	53	20	<1
377—382	"	6,89	4,65	0,17	0,21	1,87	190	21	184	265	67	20	145	49	10	<1

*Contents of Al, Fe, Mn, Ti, K, and CaCO₃ in percent; the rest of the elements in $n \cdot 10^{-4}\%$

**Oozes: CRsc — clay-radiolarian, slightly calcareous; CRt — clay-radiolarian, tuffite; CDrt — clay-diatomitic, tuffite, enriched with radiolarians; RDt — radiolarian-diatomitic, tuffite; CCr — calcareous-clay, enriched with radiolarians. See Table 1 for the rest of the abbreviations. Analysts N. N. Zavadskaya and N. P. Tolmacheva (Institute of Oceanography, Russian Academy of Sciences).

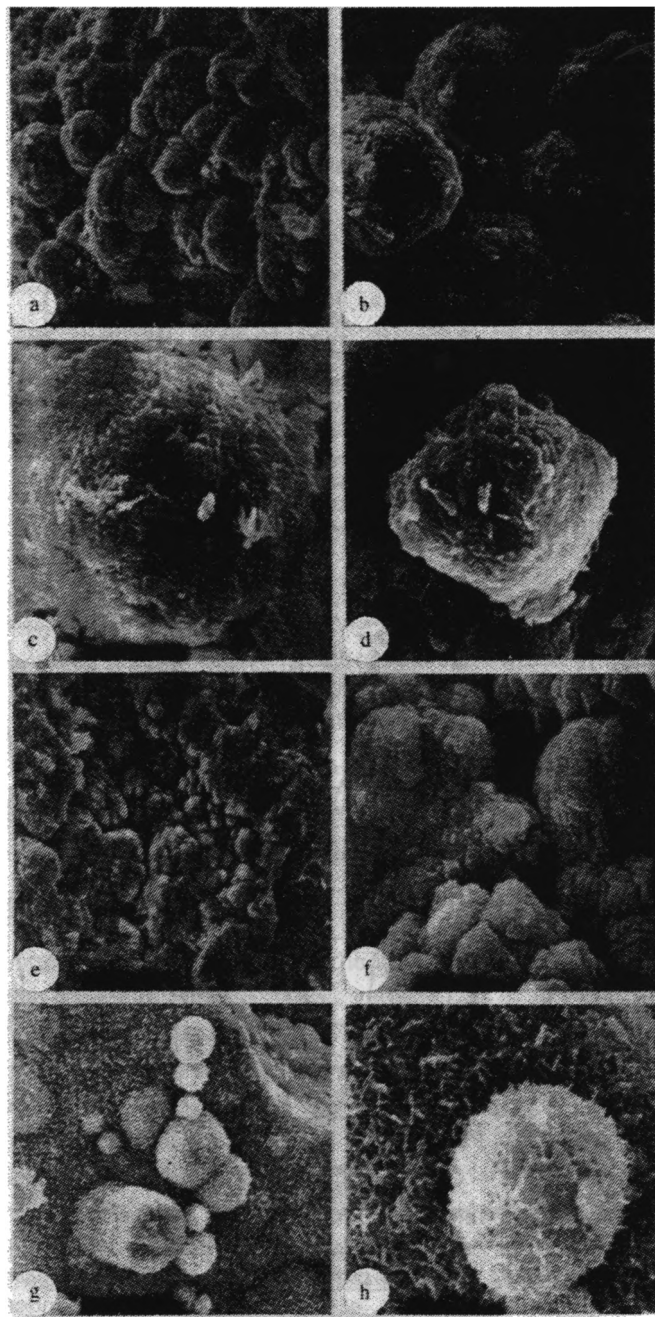


Fig. 2. Electron-microscopic photographs of calcium rhodochrosite: a-f) concretions of spheroidal rhodochrosite (a – mag. 500, b – mag. 1500, c – mag. 2500, d – mag. 2000, e – mag. 500, f – mag. 2500); g) authigenic clay minerals at the contact of rhodochrosite with decomposing ferromanganese concretion, mag. 400; h) the same, mag. 2000.

40% CaCO_3 . The opposite trend is traced in the distribution of manganese (see Fig. 1). On the whole, in relation to stations 3884 and 3899, the sediments at station 3850 are impoverished with organic matter, but enriched with manganese and calcium carbonate.

Thus, segregations of authigenic manganese carbonate in the studied columns are observed in different types of sediments, against a background of contents of residual organic carbon of 0.3-1.6%, with low carbonate content (CaCO_3 less than 1-8.3%) and concentration of manganese of 0.4-4.5%.

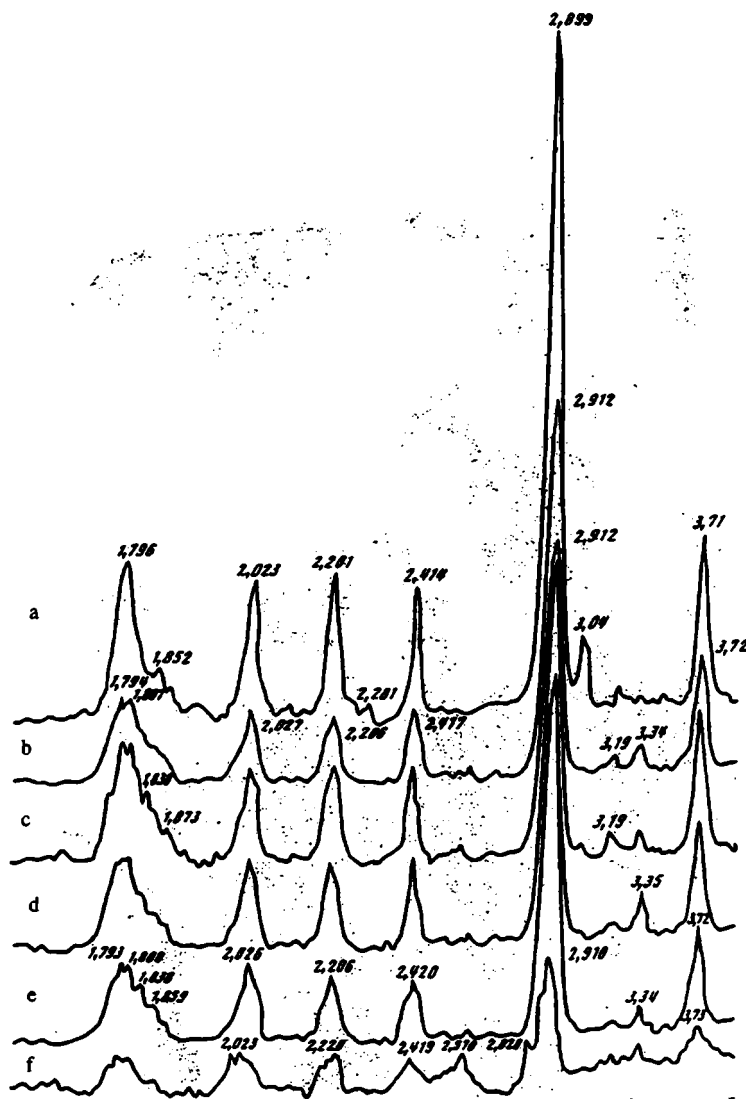


Fig. 3. X-ray diffractograms of calcium rhodochrosite: a) yellowish-white nodule, station 3850, 330-340-cm horizon; b) yellowish-gray flattened nodules, station 3884, 249-255-cm horizon; c-f) station 3899, 295-300-cm horizon (c – yellowish-gray loose nodule, d – nodule's brownish-gray loose jacket, e – nodule's gray, dense inner part, f – contact of rhodochrosite and ferromanganese concretion).

Composition of Authigenic Manganese Carbonate. Study in immersion and sections showed that authigenic carbonate is represented by spherical aggregates with N_0 1.736 and N_e 1.545. In an electron scanning microscope, concretions of spheroidal segregations of manganese carbonate are clearly traced (Fig. 2a-f). At station 3899, at its contact with a decomposing ferromanganese concretion, in a scanning electron microscope segregations of authigenic clay minerals are discovered, which sometimes probably inherit the form of biogenic remains (see Fig. 2g and h).

X-ray analysis showed (Fig. 3) that the investigated specimens are composed of rhodochrosite [21]. The excessively high basal reflection (2.899-2.912 Å) in relation to the standard pattern (reflex 2.85 Å) indicates isomorphic replacement of part of the manganese ions by calcium and magnesium ions. The purest rhodochrosite was found at station 3850 (see Fig. 3). The presence of rhodochrosite is confirmed by the results of study of the chemical composition of authigenic nodules (Table 3). After converting carbonate components to the sum of 100%, the following approximate formulas were obtained for the investigated mineral:

$(\text{Mn}_{60.7}\text{Ca}_{32.5}\text{Mg}_{6.5}\text{Fe}_{0.3})\text{CO}_3$ – loose nodule;

$(\text{Mn}_{60.4}\text{Ca}_{32.7}\text{Mg}_{6.6}\text{Fe}_{0.3})\text{CO}_3$ – nodule's loose outer jacket;

$(\text{Mn}_{64.6}\text{Ca}_{32.4}\text{Mg}_{2.8}\text{Fe}_{0.2})\text{CO}_3$ – nodule's dense inner part.

TABLE 3. Chemical Composition of Nodules of Calcium Rhodochrosite (% of dry matter) *

Component	Station 3899, 295-300 cm horizon**			Component	Station 3899, 295-300 cm horizon		
	1	2	3		1	2	3
SiO ₂	11,43	11,03	5,63	Cr ₂ O ₃	0,00	0,00	0,01
TiO ₂	0,19	0,17	0,10	NiO	0,05	0,02	0,02
Al ₂ O ₃	4,01	2,82	2,34	ZnO	0,00	0,00	0,00 -
Fe ₂ O ₃ tot	1,90	1,76	1,07	Fe ₂ O ₃	1,74	1,60	0,94
MnO tot	28,31	29,24	33,36	FeO ***	0,14	0,14	0,11
MgO ***	2,74	2,68	2,52	H ₂ O ⁺	3,07	0,58	1,79
CaO ***	13,54	13,66	15,76	H ₂ O ⁻	1,07	1,59	1,25
P ₂ O ₅	0,17	0,30	0,23	CO ₂ ***	29,20	30,00	32,50
Na ₂ O	1,22	1,30	0,60	C _{org}	3,00	5,00	3,40
K ₂ O	0,52	0,43	0,34	MnO ₂	Not found	0,21	0,17
BaO	0,06	0,01	0,03	MnO ***	28,54	29,15	33,77
SrO	0,05	0,04	0,04	Total	100,74	100,82	100,94

*Analysts V. V. Karpushina and G. A. Granovskaya (Geological Institute, Russian academy of Sciences).

**1) brownish-gray loose nodule; 2) nodule's brownish-gray loose outer jacket; 3) nodule's gray, dense inner part.

***Original components for calculating rhodochrosite's formula.

Thus, with respect to a set of traits, in sediments of the Guatamalan basin authigenic carbonate is represented by calcium rhodochrosite with a noticeable admixture of magnesium. Rhodochrosite previously discovered in this basin [22] has a close composition: (Mn₅₀₋₈₀Ca₂₀₋₅₀)CO₃. In borehole 503 (south of the Guatamalan basin), calcium rhodochrosite forms numerous nodules in reduced sediments of the Late Miocene–Holocene [16]. These authors suggest that rhodochrosite was formed near the bottom surface as a result of the interaction of bicarbonate in bottom and pore waters with bivalent manganese in a very high concentration.

Isotopic Data and Discussion of Them. The overwhelming majority of available isotopic data for contemporary manganese-carbonate formations relates mostly to lacustrine or marine sediments. Specifically oceanic manganese carbonates have not been studied much isotopically, which is due to the insufficient degree of examination of regions of the World Ocean where they are most likely to be found.

One of the known basins with widespread development of Mn carbonates in contemporary sediments is the Guatamalan basin. Isotopic data available in the literature [19, 23] relate to carbonates taken precisely in this basin.

The material and isotopic composition of carbon and oxygen in 17 specimens rich in manganese carbonates that were disclosed in boreholes 503A and 503B was studied in detail in [19]. The overall limit of variations in these samples' isotopic composition (southern part of the Guatamalan basin) was from -3.8 to -1.2‰ for $\delta^{13}\text{C}$ and from 4.06 to 5.99‰ (in relation to the PDB standard) for $\delta^{18}\text{O}$. The isotopic composition of manganese carbonates studied in the present work proved to be analogous (Table 4): the values of $\delta^{13}\text{C}$ vary in the range from -2.5 to -1.1‰; and $\delta^{18}\text{O}$, from 34.9 to 37.1‰ (in relation to the SMOW standard).

The isotopic data given differ significantly from those characteristic of lake and marine Mn carbonates in the direction of higher values of $\delta^{13}\text{C}$. At present, it is not fully possible to compare the isotopic composition of the studied rhodochrosites with analogous formations from other parts of the World Ocean. The isolated isotopic data at our disposal for other finds of oceanic manganese carbonates, for example, rhodochrosite from a column at station 82105 (15-25 cm horizon) located on the slope of the Central American trench (Elgard uplift), are also characterized by lighter isotopic composition: $\delta^{13}\text{C} = -6.0\text{‰}$ (the specimen was studied together with L. E. Shterenberg et al. [16]).

Thus, the rhodochrosites in sediments of the Guatamalan basin differ significantly in their isotopic characteristics from analogous formations, in the direction of higher contents of the heavy isotope ^{13}C . This is apparently connected with the specific conditions of formation of Mn carbonates in the sediments of this basin.

It has been established that practically all known manganese-carbonate formations (fossil and contemporary marine and lacustrine [2-6]) are characterized by low values of $\delta^{13}\text{C}$. This is due to the fact that Mn carbonates in sedimentary

TABLE 4. Isotopic Composition of Carbon and Oxygen in Sediments of the Guatamalan Basin

Analysis number	Specimen number	Place where specimen was taken and its characteristics	$\delta^{13}\text{C}$, ‰ (PDB)	$\delta^{18}\text{O}$, ‰ (SMOW)
2850	3884-250	Station 3884, 249-253 cm horizon. Yellowish-gray carbonate nodules: calcite, rhodochrosite	-1,9 -1,6	35,2 35,5
2851	3899-1	Station 3899, 295-300 cm horizon. Brownish-gray loose carbonate nodules: calcite, rhodochrosite	-2,6 -2,3	35,7 34,9
2852	3899-2	The same place. Brownish-gray loose jacket of carbonate nodule: calcite, rhodochrosite	-2,6 -2,3	35,3 36,9
2853	3899-3	The same place. Gray dense inner part of carbonate nodule: calcite, rhodochrosite	-2,3 -1,9	35,4 36,6
2854	3899-295	The same place. Yellowish-gray carbonate nodule with biogenic remains: calcite, rhodochrosite	-3,0 -2,5	32,1 36,0
2855	3899-301-306	The same place, 301-306 cm horizon. Greenish dark-gray clay-diatomic ooze enriched with radiolarians: rhodochrosite	-1,9	37,1
2856	3850-335	Station 3850, 330-340 cm horizon. Yellowish-white carbonate nodule: calcite, rhodochrosite	0 -1,1	33,3 36,3
2857	3899	Station 3899, 295-300 cm horizon. Sediment washed off of nodules: rhodochrosite	-2,3	—
2858	3899	The same place. Carbonate nodule: calcite, rhodochrosite	-0,9 -1,0	— —
2859	3899	The same place, 300-302 cm horizon. Sediment rinsed with water: rhodochrosite	-2,0	—

strata are formed in the stage of diagenesis with the active participation of isotopically light carbon dioxide. The latter forms as a result of oxidation of organic matter within the sediment itself. Therefore, we can suppose that formation of Mn carbonates in sediments of the Guatamalan basin as a result of diagenesis occurred in somewhat different, fairly specific conditions differing in that the role of CO_2 of organic origin in this process was insignificant. The formation of manganese carbonates occurred mainly as a result of dissolution and redeposition of biogenic calcium carbonate in the sediments (apparently the shells of foraminifers and calcareous skeletal remains of other organisms).

For some of the samples at our disposal, we were able to study different genetic varieties of manganese carbonates contained in the same sample. Thus, the isotopic composition of carbon in rhodochrosite from the surrounding sediment and from a carbonate nodule in the same interval of column 3899 proved to be identical: $\delta^{13}\text{C}$ varies in the range of -2.5 to -2.0 ‰ (analyses 2851, 2852, 2857, 2859). In this case, in some nodules their inner part is distinguished by higher values of $\delta^{13}\text{C}$ (-1.9 ‰, analysis 2853) in comparison with the outer jacket (-2.3 ‰, analysis 2852). Analogous results were obtained previously in [19]: for two specimens out of five studied, the isotopic composition of carbon gradually becomes heavier from the periphery toward the center.

On the whole, the isotopic composition of manganese-carbonate material does not depend on the form in which it is found in the sediment (nodules, dispersed, etc.). The slight observed variations in values of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ in different samples, as well as within the bounds of a single nodule, are most likely connected (as is suggested in [19]) with local conditions (the sediment's permeability, physicochemical inhomogeneity, etc.) of formation of Mn carbonates.

TABLE 5. Isotopic Composition of Carbon and Oxygen in Foraminifers from Sediments of the Guatamalan Basin

Analysis number	Specimen number	Place where specimen was taken	$\delta^{13}\text{C}$, ‰ (PDB)	$\delta^{18}\text{O}$, ‰ (SMOW)
Foraminifers (>0.25 mm) from sediment				
3350	3850	Station 3850, 330-340 cm horizon	2,0	29,9
3351	3884	Station 3884, 248-253 cm horizon	1,0	30,0
3352	3899	Station 3899, 295-300 cm horizon	0,5	31,2
Formation from nodules				
3353	3850	Station 3850, 335 cm horizon	0,6	32,8
3354	3884	Station 3884, 250 cm horizon	-1,1	25,1

For comparison, we studied the isotopic composition of carbon and oxygen in foraminifers, which were taken from the surrounding sediment, as well as from manganese-carbonate nodules themselves (Table 5). Here, a consistent difference is displayed in values of the isotopic composition of foraminifers and manganese carbonates from nodules. On the whole, the former are characterized by higher values of $\delta^{13}\text{C}$ (by approximately 2‰) in relation to the nodules. This difference is apparently due to different isotopic compositions of the original bicarbonate and the temperature of formation of carbonate material. It is well known that the bicarbonate in upper horizons of the hydrosphere is characterized, as a rule, by higher values of $\delta^{13}\text{C}$, which is a consequence of the higher bioproductivity of surface horizons of the water mass. Therefore, the carbonate in foraminifers, which was formed on account of the bicarbonate in upper horizons of the hydrosphere, should have higher values of $\delta^{13}\text{C}$.

The isotopic composition of oxygen in foraminifers in comparison with Mn carbonates in the surrounding sediment, on the other hand, is slightly enriched with the light isotope ^{16}O and, on the whole, is characterized by lower values of $\delta^{18}\text{O}$ (29-33‰) corresponding to the equilibrium for carbonates deposited from surface horizons of seawater.

* * *

1. In the Guatamalan basin, authigenic manganese carbonate represented by calcium rhodochrosite is widespread. Nodules and microsegregations of this mineral gravitate toward different-aged reduced and slightly oxidized terrigenous-biogenic ooze in the transition zone from hemipelagic sedimentation conditions to miopelagic.

2. Along with pyrite, rhodochrosite can serve as a reliable criterion for separating regions of continental-margin and pelagic types of lithogenesis in the ocean.

3. Calcium rhodochrosite in the studied columns was formed against a background of contents of residual organic carbon in the sediments of 0.3-1.6%, with their low carbonate content (<1-8.3% CaCO_3) and manganese concentration of 0.4-4.5%.

4. The carbonate material in the studied rhodochrosites was taken from two sources: sedimentary calcium carbonate and isotopically light CO_2 of organic origin. In this case, the former is predominant and is formed as a result of dissolution and redeposition of calcareous shells of foraminifers and skeletal remains of other organisms. Isotopically light carbon dioxide is formed as a consequence of oxidation of C_{org} and plays a subordinate role in the process of formation of authigenic carbonates.

5. The source of Mn in the studied nodules was the surrounding sediments, and sometimes Fe-Mn concretions, which supplied Mn as a result of its remobilization in the process of diagenesis.

6. The main feature of formation of Mn carbonates in sediments of the Guatamalan basin is the low geochemical activity of organic matter in this process, which resulted in relatively high values of $\delta^{13}\text{C}$ in the nodules.

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CARBONATES IN LAKES OF THE NORTHERN PART OF THE RUSSIAN PLATFORM

L. E. Shternenberg

UDC 552.54:551.312(470.3)

In the article, we present results of a comprehensive investigation of sedimentation features in lakes of the northern part of the Russian platform (the Leningrad Oblast, Latvia, Lithuania, and Estonia). We show that deposition of carbonates in these water bodies primarily takes place via the terrigenous pathway.

Questions related to the formation of carbonates in recent inland bodies of water in the northern humid zone have been the subject of a number of investigations, including especially important reports by Baranov [4], Bartosh [5-7], Garunkshtis [8], Danilans [9-11], and Lukashev et al. [13-16], among others.

Strakhov and his coworkers [24-28] have made major contributions to the subject of carbonate deposition in water bodies of various climatic zones as well as in water bodies of various dimensions and types (lacustrine, marine, and oceanic). Strakhov gave special attention to basins of the humid zone. He distinguished between two essentially different types of carbonate deposition in such basins. The first is associated with fresh water inland seas and inlets as well as lakes containing soft water; the second is associated with lakes containing hard water, typically disseminated among water bodies of the first group. Within continents, such water bodies are often restricted to a transitional belt extending from a zone of humid climate to a semiarid zone, where they are more typical.

In Strakhov's opinion, appreciable undersaturations of solutions with respect to calcium carbonate salt are created in fresh-water northern lakes with soft water, small alkali reserves (0.5-1.5 mg/liter), and large quantities of corrosive CO_2 .

An organic system capable of precipitating carbonates is very weakly developed (some species of gastropods and bivalves), but shelly remains left following the death of an animal are almost completely dissolved by water and the sediment stripped of its carbonate component. Examples of this include such lakes as Ladozhsk and Onezhsk, etc.

Carbonate deposition in lakes with hard water, rich in Ca and Mg, is quite different. These lakes are typically located within areas where carbonate rocks also occur. Carbonate deposition is most typical and clearly manifest in lakes situated along the southern margins of humid taiga-podzolic zones and in forest-steppe belts (lakes in northeastern states of the USA and Western Europe, including Germany, Denmark, and Sweden, as well as in the Commonwealth, primarily in Byelorussia, the northern Ukraine, and the Trans-Urals).

Favorable conditions for carbonate formation in such bodies of water exist in the epilimnion, approximately corresponding to the littoral and sublittoral zones of the near-shore zone. During summer stagnation (maintained by vertical zoning of water) and occasionally year-round stagnation, water of the epilimnion becomes over-saturated with CaCO_3 and, consequently, may promote chemical precipitation of CaCO_3 . Lime-secreting phytobenthos and zoobenthos abound here. The least favorable conditions for accumulation of CaCO_3 are created in the profundal (hypolimnion). Plant and animal organisms are almost completely absent in these parts of lakes. The water of the hypolimnion become abundant with free CO_2 over the course of an annual cycle and promotes active dissolution of CaCO_3 arriving to the depths by any path.

Thus, hard-water lakes of the humid zone, according to Strakhov, are characterized by chemical precipitation of calcium carbonate from water, biogenic settling by means of lime-secreting blue-green algae and, finally, biogenically precipitated calcite (CaCO_3) in the form of shells of organisms.

In addition, different hypotheses concerning the conditions of formation of carbonates in fresh water bodies in the northern part of the Russian platform have been proposed by a number of investigators studying sediment deposition in lakes of the Leningrad Oblast, Byelorussia, and the Baltic region (Estonia, Latvia, and Lithuania, etc).

Rossolimo [22] has proposed that calcium-depositing types of lakes are characterized by sedimentation of calcium in the carbonate form, with calcium supplied to the lake as hydrocarbonates, predominantly from groundwater. Deposition of calcium takes on the significance of a typological indicator when the presence of calcium-containing substances, suggesting a predominance of processes stabilizing this element, are discovered in recent bottom deposits of lakes.

Garunkshtis [8] believes that, in lakes of Lithuania, the chemical composition of the water filling a lake determine the processes of sediment (including carbonate sediment) deposition. In his opinion, precipitation of hydrotroilite and calcium carbonate from bottom water is governed by the quantity of soluble components remaining in the water. Garunkshtis does not take into account nor discuss suspended matter entering a lake and also "forgets" processes taking place in the sediments themselves during their accumulation and transformation within the stage of diagenesis.

In one of his works, Bartosh [5] discusses geological aspects and resources of fresh water carbonate distributions of the Holocene in the central zone of the European part of the USSR. Although the investigations by Strakhov, in the opinion of Bartosh, allow us to construct a general scheme of carbonate formation in recent bodies of water, there are "... a number of questions concerning carbonate accumulation which are still unresolved and there are still different opinions concerning the governing factors promoting or impeding formation of carbonates and their precipitation from oversaturated bicarbonate water" [5, p. 20].

Considering the climatic features associated with carbonate deposition, it seems to us that Bartosh has not taken into consideration the important data offered by Strakhov concerning details of this process in regions located in northern humid zones where the catchment basins primarily consist of carbonate-bearing rocks and where occurrences of such formations in lakes are not at all associated with aridization of climate.

Formation of calcareous (calcitic) deposits, according to Bartosh, takes place as a result of supply of groundwater rich in HCO_3^- and Ca to a lake. However, the data he presents does not indicate such a formation pathway for recent carbonate sediments in lakes. The author neglects the diluting action of lake waters, impeding many-fold oversaturation of components belonging to a carbonate system. Alekin and Moricheva [2] as well as other investigators have demonstrated the necessity of such oversaturation. Structural positions of such deposits (the presence of carbonate sediments at two sides of a lake basin), in our opinion, also contradicts Bartosh's suggestions. The claims concerning of calcite precipitation are not very precise. We have recalculated the results of chemical analyses of carbonate presented in Bartosh's reports [5-7] and have shown that, in addition to CaCO_3 , MgCO_3 is present in quantities reaching 14-15% or more in individual cases. Bartosh probably did not consider the possibility that lacustrine carbonate deposits may form as a result of their overland supply from catchment areas, neither did he take into account the participation of algae in these processes.

Lukashev et al. [13-16] assert that accumulation of carbonate sediments in the lakes of Byelorussia takes place only in the litoral and sublitoral zone in the case of year-round oversaturation of water with carbonates and an absence of corrosive CO_2 . Their data suggests that only CaCO_3 is chemically precipitated. Very soluble forms of magnesium present in lacustrine sediments and passing into a slightly acidic extract exclusively bind to terrigenous dolomite. In the opinion of these investigators, carbonate formation of this type in lakes of the Baltic region and Poland to the west of Byelorussia is still dominated by sediment deposition at this time. An important reason for this is the high carbonate content of the glacial sediments which enclose the basins of recent bodies of water.

Nikolayeva [20] has discussed CaCO_3 deposition in sediments of lakes in the former Mariysk and Tatarsk ASSR, i.e., in regions where karst processes and water catchment areas substantially composed of carbonate rocks are widespread. This report indicates that despite the high overall hardness of water supplied from springs, streams, etc., clear deficits of CaCO_3 can be detected in the hypolimnions of many lakes. Only 2 lakes out of 15 show positive values for the difference between CaCO_3 found and computed from the solubility of CaCO_3 . It is interesting to note that, even in these lakes, Nikolayeva was not able to find clearly chemogenic calcite. He found only former shells and terrigenous-carbonate muds in these lakes. On the other hand, he found small limestone crusts and incrustations on growths of underwater vegetation in the near-shore zones of other lakes where, according to computational data, carbonate precipitation is impossible.

As we can see, the evidence presented above suggests that there are various plausible hypotheses concerning carbonate formation on lake bottoms in the northern part of the Russian platform. Most likely, this is the result of the fact that the authors of the majority of reports mentioned above, when addressing the problem of carbonate deposition in lakes, rest their conclusions primarily on chemical determination of water and sediment compositions, forgetting the necessity of a comprehensive investigation of such features for a resolution of this complex problem.

Methods and Objects of Investigation. In contrast to the preceding reports based on investigations of sediment and carbonate formation in lakes of the northern part of the Russian platform, we carried out a comprehensive investigation of a number of lakes in the Leningrad Oblast, Latvia, Lithuania, and Estonia.

Sediments and water samples from the lakes were collected in a specially equipped sailboat of the "Meva" type to which a outboard engine was attached at the transom. The boat had bulging sides and therefore was very stable in water. Sediment collection was carried out by means of a Perfil'ev-type stratometer. Bottom and surface waters were collected by means of a bathometer.

Water analyses were carried out in two phases: in a field phase and in an office phase. In the field, abundances of various forms of iron, manganese, CO_2 , HCO_3^- , C_{org} , SO_4^{2-} , and oxygen, as well as permanganate oxidizability were determined; pH, Eh, temperature, etc. were measured; initial descriptions of sediments brought up by the stratometer was carried out. In the office phase, water samples were analyzed for abundances of magnesium, calcium, and amorphous silica using Sochevanova's method [23]. Mechanical analysis (aqueous sieving and elutriation) of sediments, and also mineralogical investigation (macroscopic and microscopic examination, X-ray analysis with the aid of a diffractometer and Debaye-Sherer chamber, thermographic analysis, staining of thin and polished sections, etc.) were carried out by the authors. Contents of Fe, Mn, C_{org} , CO_2 , and other components in sediments were determined in the chemical laboratory of the Institute of Geological Sciences of the Academy of Sciences of the USSR.

The geological makeup of the study area includes deposits of the Jurassic, Triassic, Permian, (light-gray limestones), Devonian, represented by terrigenous and lagoonal sediments (a sandy-clayey complex with interbeds of sand, clay, marl, and gypsum), Ordovician, and Silurian (sandstone, siltstone, limestones, and marl).

A Quaternary deposits overlying the mantle of older rocks includes moraine clay, sandy loam, loam with gneiss and granite boulders, and fluvio-glacial and lacustrine-glacial deposits.

Direct proximity to the Baltic Sea has had a substantial influence on the climatic classification of the territories of Estonia, Latvia, and Lithuania, which show transitional features between marine and continental conditions. A moderate temperature regime (a yearly average temperature of 4.2-6.9°C.) elevated humidity (140-195 days per year with relative humidity of $\approx 80\%$), and elevated precipitation (650-950 mm/yr) predominate in this region. In the Baltic region, sod, podzolic, swampy, and moor types of soil predominate, although podzolic-gley soils occur in individual areas. The region is classified as a mixed-forest zone (pine is most abundant). In the Baltic region, marshes and peat bogs (low, transitional, and high peat bogs) are widespread.

The present-day drainage system belongs to the basin of the Baltic Sea and consists of numerous rivers which criss-cross the territory rather densely. Calcium and magnesium hydrocarbonates are the main constituents of the water.

The territory under discussion is very abundant with small lakes (for example, there are ≈ 1150 in Estonia and more than 3000 in Latvia). The overwhelming majority are very small ($< 0.5 \text{ km}^2$ of water surface), and only a very small number (42) can be classified as intermediate in size, with areas of ≈ 100 hectares.

Within the territory, we investigated 16 lakes (7 in Estonia, 4 in Latvia, 2 in Lithuania, and 3 in the Leningrad Oblast). These lakes differ from one another in size, depth, morphology, and other indices. Published data and data we have collected suggest that the Baltic region can be described as showing generally the same character of sediment deposition in lakes; this is most likely a result of the fact that the area has the same geomorphological and geological structures in common, a single climatic zonation, similar processes of soil formation, and other factors.

This territory is located in the northwest part of the East European plain, where elevations below 120 m predominate. The region is mainly one of small hills and is characterized by frequent alternation of positive and negative terrain forms associated with glacial deposits.

Myaemets [19] has presented an extensive hydrobiological description of the lakes of Estonia. Myaemets identifies 12 types of water bodies differing in transparency, total mineralization of water, pH, HCO_3^- ion content, bichromate oxidizability, abundance of chlorine, depth, and a number of other indices. Judging from these data, we might assume that very diverse types of sediments have accumulated in the lakes of Estonia as well as in the lakes of Latvia and Lithuania. However, extensive published data and our own observations allow us to substantially narrow the classification of lakes in the Baltic region.

Most likely, Gorunkshtis [8] is correct to distinguish only three principal types of deposits (organogenic, terrigenous, and carbonate) accumulating in recent water bodies.

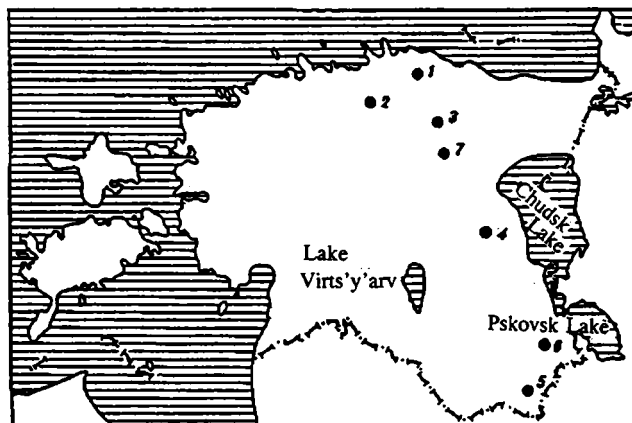


Fig. 1. Map showing locations of lakes investigated in Estonia: 1) Viitna; 2) Purgateli; 3) Parkuni; 4) Caad"yarv; 5) Kirikumyae; 6) Must"yarv; 7) Yantu Siniyarv.

Based on the results of preceding investigations and our own, we can assert that similar deposition of sediments takes place in lakes of the Baltic region (Estonia, Latvia, and Lithuania). Here, we believe it is appropriate to dwell on a description of lakes in Estonia, and to generalize the data obtained to bodies of water in Latvia and Lithuania.

A schematic map showing locations of Estonian lakes we investigated is shown in Fig. 1. Published data and, to a substantial degree, the results of investigations carried out by Myaemets [19] allow us to categorize the lakes investigated into three groups: 1) Kirikumyae, Viitna, Must"yarv, Purgateli; 2) Yantu Siniyarv; 3) Caad"yarv, Parkuni.

Results Obtained. First Group of Lakes. All bodies of water in this group are small and shallow. Typically, the coasts of such lakes are low-lying, frequently marshy. The water has a low level of mineralization (9-13 mg/liter). According to the classification of Alekin [1], the water is of the sodium-sulfate type. Saturation of surface water with oxygen does not reach 100% even in summer; it varies from 67 to 90%. The permanganate oxidizability of these waters is great, although it reaches 36-37 mg/liter O_2 /liter in the very marshy Lake Mus"yarv. Results of determinations of abundances of Mn, as well as various forms of iron, silica, CO_2 , and HCO_3^- in the lake waters are presented in Table 1. These data indicate that carbon dioxide is present in all horizons and increases to 16-32 mg/liter toward the lake bottom. The content of HCO_3^- in the water is not very appreciable (up to 9 mg/liter). We did not find manganese in these lakes. Among various forms of iron, the most important is iron bound with organic acids in complex coordination compounds. Sandy sediments are found in all of the investigated lakes of this group, but they occur differently in different lakes. In Lake Kirikunya, they form small lenses and strips restricted to the western lakeside. In terms of granulometric composition, they can be classified as fine-grained, somewhat silted varieties. At Lake Must"yarv (8.9-m deep in its deepest part), arenaceous sediments almost completely rim the narrow strip of shore. On the eastern shore, these sediments form beaches. Sands are represented by various types. The largest-sized are coarse-grained varieties. In the fine fraction of sands (0.1-0.25 mm) from lakes of the first group, quartz clearly predominates (76-82%). Plagioclases account for approximately 13%, and K-feldspars average 10%. There are isolated occurrences of micas (muscovite and biotite) and carbonate fragments. Peaty-sapropelic muds are found in the centers of lakes Kirikumyae, Must"yarv, Viitna, and Purgateli. Under a microscope, it is evident that fairly well preserved plant remains are components of these deposits. The groundmass is probably of an allochthonous humic origin, although we cannot rule out the participation of sapropelic material formed in situ in the lakes. Such dystrophic and eutrophied bodies of soft water, as a number of reports have demonstrated, are rich in desmid algae. The mineral admixture is fine-grained. Grains are dispersed throughout the organic mass and do not form accumulations. Contents of Fe, Mn, and SiO_2 in sediments of lakes Kirikumyae, Viitna, Must"yarv, are Purgateli are very low. In contrast, contents of C_{org} are distinctly elevated (40-41%) (Table 2).

The data we obtained suggest accumulation of peaty-sapropelic muds and peaty-sapropels in these lakes.

Lake Yantu Siniyarv belongs to the second group. According to Yaemets [19] these bodies of water are of the alkali-trophic type. Lakes of this group occur in Northern Estonia, in the Pandive uplands. They are Precambrian fed by groundwater and spring water. High contents of HCO_3^- (>3.4 mg-equiv./liter) and low amounts of resistant organic

TABLE 1. Abundances of Biogenic Components in Waters of Estonian Lakes (summer of 1971), mg/liter

Lake (station)	depth of water-sample collection, m	pH	Eh, mV	CO ₂	HCO ₃ ⁻	Fe ²⁺ + Fe ³⁺	SiO ₂
Kirikumäe	0.20	5.2	+380	4.95	0	Traces	—
	2.30	6.3	+340	2.30	4.88	0.3	
Viitna	0.20	6.8	+410	1.40	2.68	0.1	2.5
	6.00	6.6	+395	2.45	3.66	0.1	3
Must"yarv	0.20	4.2	+490	14.00	0	0.4	2
	4.85	—	—	16.50	0	0.4	2.5
Purgateli	0.20	—	—	2.10	3.66	0.05	2
	8.20	—	—	32.27	9.76	3	
Caad"yarv	0.20	8.20	—	0	183.0	0.1	2.5
	14.10	7.90	—	1.65	193.0	0.1	4.0
Parkuni	0.20	8.35	+400	2.45	222.0	Traces	3
	0.40	8.20	+370	9.90	231.8	0.1	4
Yantu Siniyarv [1]	0.20	8.00	+405	2.30	270.8	0.05	4
	4.80	7.70	+400	2.30	268.4	0.05	4

Explanation. In this and the following tables, lines across columns indicate lack of data. Fe²⁺ and Mn were not detected. Data from the analysis of samples collected from the surface of the water are shown in the numerator; data from the samples collected from the near-bottom portion are shown in the denominator.

matter (permanganate oxidizability < 7 mg O₂/liter) are typical. According to the data of a number of Estonian investigators, lake marl is accumulating in Lake Yantu Siniyarv. Plankton productivity is low, since the large quantity of calcium in the water to some degree impedes migration of phosphorus and iron ions.

Lake Yantu Siniyarv is a small body of water 1.2 km² in area. It consists of three pools of water which are joined to one another. According to our measurements, the lake reaches a depth of 5 m. The water is very clear (the bottom is visible) and light-green in color. The pH of the water equals 8.0 in the surface part and 7.7 in the bottom part. The content HCO₃⁻, the principal anion, is 270-268 mg/liter. Manganese has not been found in the water. Among forms of iron, iron bound with organic matter in coordination compounds predominate. The amount of silica in the water is approximately the same in the surface zone and at the bottom, equalling 4 mg/liter (see Table 1).

Mud, which is abundant over the entire aquatory of the lake, is light-gray in color. It is very soft and is easily kneaded by the fingers after desiccation. Carbonate crusts up to 1-1.5 cm thick, surrounding the branches and stems of large shrubs, were raised up from the bottom in the nearshore zone. According to determinations by N. P. Gorbunova (Lomonosov State University) these crusts owe their origin to blue-green algae (*Microcoleus* and *Colothrix*).

Chemical analyses of crusts and mud reveals insoluble mineral residue differing only in quantity (Table 3). The quantity of insoluble mineral residue equals 8% in mud and 1.5% in crusts. Let us now turn our attention to the especially similar abundances of carbonates and organic matter in these samples (C_{org} of the crusts equals 4.3%; C_{org} of the mud equals 3.4%); this suggests similar origins. According to the analyses, MgO, in addition to CaO, is present in the carbonate muds and crust.

Heating curves (results of thermographic analysis) of mud crust from Lake Yantu Siniyarv are presented in Fig. 2. Analysis of the crust (see Fig. 2b) reveals the presence of a single carbonate mineral (an endothermal effect at 880°C) and a large exothermal maximum associated with the decomposition of algal material divided in two by a small endothermal interruption at 440°C. On the heating curve for carbonate mud from Lake Yantu Siniyarv (Fig. 2a), the shift of the main endothermal peak toward the lower-temperature region in comparison with the heating curve for the crust is most likely due to the influence of a clay admixture [29].

Although the X-ray diffractometer curve of the crust, like the heating curve, suggests the presence of calcite (3.03 Å), a certain asymmetry of the side of the peak, facing the region of higher angles, allows us to assume the presence of yet another carbonate mineral here as an additional phase. The presence of two minerals differing in composition, calcite (predominating) and a low-magnesian calcite, was confirmed by staining of thin-polished sections [30]. As a verification, the material composing the crust was separated in a heavy liquid, M-45, by centrifugation into light (< 2.75) and heavy

TABLE 2. Abundances of Components in Sediments of Investigated Estonian Lakes, %

Lake, station no.	Fe	Mg	P	C	CO ₂	SiO ₂ (amorph.)	Type of deposit
Kirikumyae							
Station 1	0,50	None	0,02	41,37	0,45	4,55	Peaty sapropel
Station 2	0,39	"	0,02	None	None	—	Fine-grained sand
Viitna							
Station 1	2,12	0,04	0,03	0,21	"	—	Intermediate-grained sand
Station 2	4,08	0,03	0,05	29,45	1,60	1,05	Peaty-sapropelic mud
Must"yarv							
Station 3	1,09	0,02	0,02	0,71	1,20	1,12	Coarse-grained sand
Station 2	0,59	Her	0,04	40,46	None	4,20	Peaty sapropel
Purgateli							
Station 1	0,75	0,10	0,13	33,39	"	—	Peaty-sapropelic mud
Station 2	2,00	0,02	0,05	1,02	"	0,75	Fine-grained sand
Caad"yarv							
Station 2, sample 1	1,15	0,06	0,06	5,64	18,45	—	Clayey-limey mud
Station 3	0,47	0,03	Trace	3,74	37,55	—	Limey mud
Station 4	0,91	0,04	0,05	24,54	16,54	—	Sapropelic clayey-limey mud
Station 5	0,39	0,01	Trace	1,81	38,05	2,05	Limey mud
Station 7, Sample 1	0,67	0,02	"	5,02	34,35	—	"
Sample 2	1,88	0,03	"	4,88	35,00	—	"
Parkuni							
Station 1, Sample 1	0,20	None	0,003	1,14	40,85	0,70	"
Sample 2	0,54	"	0,003	2,82	38,80	—	"
Sample 2a	0,43	"	0,04	11,42	12,55	—	Slightly sapropelic clayey-limey mud
Station 2	0,99	0,01	0,10	16,04	0,35	1,30	Slightly sapropelic clayey mud

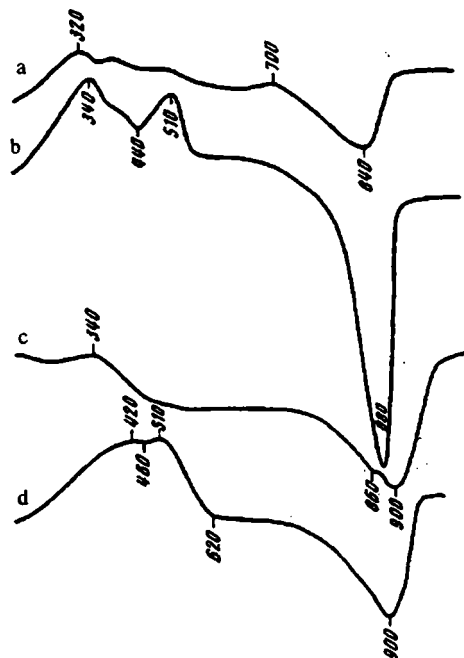


Fig. 2

Fig. 2. Heating curves of mud and carbonate crusts from Lake Yantu Siniyarv: a) mud; b) crust; c) fraction > 2.75 , crust; d) fraction < 2.75 , crust.

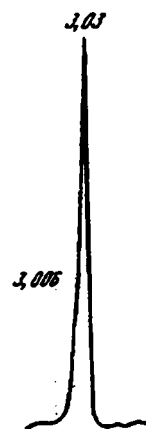


Fig. 3

Fig. 3. Diffractometer curve of carbonate crust: fraction < 2.75 , Å.

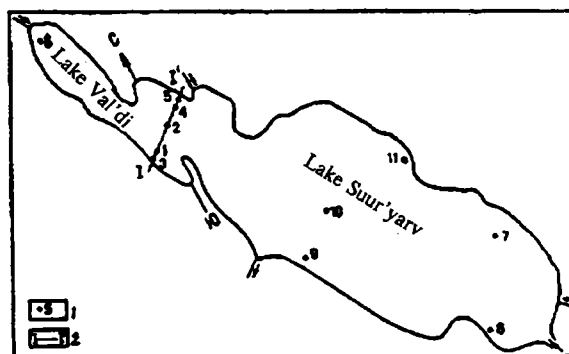


Fig. 4. Locations of sampling stations in Lake Saad'yarv: 1) Station no.; 2) lithological profile.

(> 2.75) fractions. On the heating curve of the heavy fraction (see Fig. 2c) two endothermal interruptions are evident, indicating the presence of two phases: calcite and low-magnesian calcite. In the light fraction, there is an endothermal effect at 900°C , corresponding to calcite (see Fig. 2d).

On the diffraction pattern of the light fraction (< 2.75) of crust (Fig. 3), only one peak with d 3.03 Å can be traced. This peak is somewhat asymmetric. Its side is turned toward the region of higher angles and is somewhat skewed in its lower part. A measurement at the point where the side of the peak bends suggests the presence of low-magnesian calcite (d 3.006 Å) as a minor admixture. In the heavy (> 2.75) fraction of carbonate crust, which is lesser in quantity than the light fraction, X-ray analysis also indicates the presence of two minerals.

TABLE 3. Chemical Compositions of Carbonate Formations of Lake Yantu Siniyarv, %

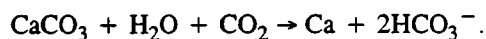
Type of sediment	C _{org}	Insoluble mineral residue	P ₂ O ₅	FeO	MnO	CaO	MgO	CO ₂	Results of tentative conversion to salts		Surplus components remaining after conversion to salts
									CaCO ₃	MgCO ₃	
Limey mud	3,4	8,04	None	0,22	None	44,51	0,62	35,80	79,65	1,30	CO ₂ 0,18
Calcareous crust on bushes	4,32	1,15	0,32	»	»	44,38	0,42	35,15	79,22	0,88	MgO 0,14

TABLE 4. Abundances of Components in Water from Different Collecting Points within Lake Yantu Siniyarv, mg/liter

Type of water	Ca ²⁺	Mg ²⁺	HCO ₃ ⁻	CO ₂
Surface (center of lake)	38,2	16,9	270,8	2,30
Bottom (same)	44,9	13,2	268,4	2,30
Bottom, near shore where crusts occur	37,4	15,2	269,3	2,30

The great similarity, as indicated by a number of indices, between the carbonate crusts on bushes and the light-gray mud filling of the bottom of Lake Yantu Siniyarv, allows us to assume that their formations were similar. Results of dissolution of light-gray mud in 1 % acetic acid of also suggestive of this. Remains of blue-green algae were found in both mud and crusts after dissolution.

As is well known, waters of lakes located at temperate latitudes are of the hydrocarbonate type and are subject to fairly complex seasonal changes accompanied by conversion of calcium and magnesium into carbonate salts and their dissolution back again. The solubility of calcium carbonate is substantially less than that of hydrocarbonate or calcium sulfate. Precipitation of CaCO₃ from water, like dissolution of this salt, is governed by the following equation:



Kuznetsov [12] proposes that precipitation of CaCO₃ from solution is regulated by the physical-chemical conditions associated with equilibrium of Ca²⁺, HCO₃⁻ ions and molecular carbon dioxide CO₂ in water, and that the process can be expressed in terms of Kol'tgof's formula

$$K = \frac{\text{Ca}^{2+} \text{HCO}_3^- \alpha}{\text{CO}_2} = 1,13 \cdot 10^{-4},$$

where Ca²⁺ is the concentration of calcium ions, in g · eq/liter; HCO₃⁻ is the concentration of HCO₃⁻ ions, g · eq/liter; CO₂ is the concentration of carbon dioxide, g · mole/liter; α is ionic activity (a quantity near unity); *K* is the equilibrium constant. According to Kuznetsov and other investigators, if the constant *K* found experimentally is less than the equilibrium constant, precipitation of calcium carbonate will not take place and calcium carbonate will dissolve. If *K* is greater, then it can be assumed that the solution is oversaturated relative to the equilibrium state of CaCO₃ and precipitation of CaCO₃ is possible so long as the equilibrium value *K* = 1.13 · 10⁻⁴ has not been established.

We determined the contents of Ca, Mg, HCO₃⁻, and CO₂ in water of Lake Yantu Siniyarv at a number of sites, for the most part near bushes, where carbonate crusts are presently forming, and also in surface and bottom waters (Table 4).

Values of *K* were calculated using the Kolt'gof formula for all sites on the lake where water samples were collected. In all cases, the values of *K* prove fairly similar to one another, concentrating in the interval 6.0-6.3 · 10⁻⁶. This is two and a half orders of magnitude less than *K* (1.13 · 10⁻⁴). Keeping in mind that this calculation is fairly approximate, since we assumed the ionic activity to be unity, all the results obtained indicate that chemogenic precipitation of calcite from water is impossible in Lake Yantu Siniyarv. Most likely, carbonate formation here, as in other lakes of the Baltic region where similar conditions hold (carbonate rocks covered by peaty soils in the catchment areas, some small supply of mineral and organic components to lakes from surface streams, etc.), is a result of the presence of blue-green algae, rather than groundwater, as proposed by Myaemets [19] and a number of the investigators mentioned above.

The clearest example of a water body of the third group, with hard water and mesotrophic features [19], is Lake Saad"yarv, located 16 km north of Tartu. This is the largest lake in the drumlin uplands, which are covered loamy moraine. It extends from northwest to southeast, in conformity with the orientation of the drumlins (Fig. 4). The length of the lake is 6 km, and the greatest width is 1.8 km. Its maximum depth is 25 m. The area covered by water equals 7.07 km², and the area of the catchment basin is 25.1 km². The thickness of sediments deposited during postglacial times reaches 6 m; this corresponds to a sedimentation rate of 1.3 mm/yr. Many investigators have reported the same or similar sedimentation rates in lakes. Lake Saad"yarv can be tentatively subdivided into two parts: a southeastern part, Suur"yarv (Bol'shoye Lake) and a northwestern part, Val'di. The shore of Bol'shoye Lake is sandy-gravelly, with occasional boul-

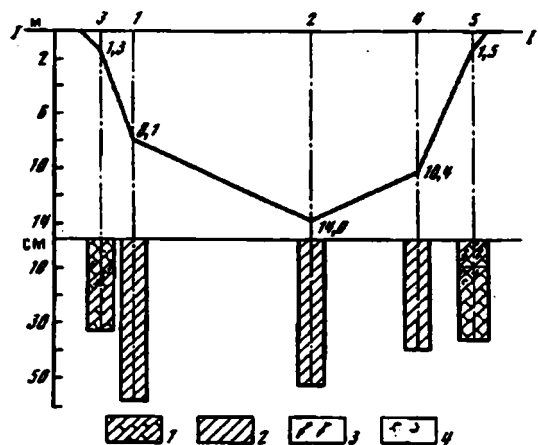


Fig. 5. Lithological profile through Lake Saad'yarv along the line I—I': 1) limey mud; 2) clayey-limey mud, sapropelic in places; 3) plant remains; 4) mollusk shells.

ders and peaty areas. The shore of the northwestern part (Lake Bal'di) is low lying and covered with shrubs. In the nearshore zone of Bol'shoye Lake, there are arenaceous sediments ranging up to coarse- and very coarse-grained sands with gravel occurring in the form of narrow, discontinuous strips. The lake water is greenish-yellow to yellowish-green. In contrast to the first group of lakes, this lake is highly mineralized (HCO_3^- up to 200 mg/liter); it belongs to the calcium-magnesium-hydrocarbonate type (see Table 1). The lake water contains relatively little organic matter, with autochthonous material predominating. Bichromate oxidizability equals 5.4 mg O_2 /liter. The percent saturation with oxygen equals 87-96% in the surface water and a total of 20% at depths of 21-22 m. The pH drops (from 8.4 to 7.5) and CO_2 starts to appear (up to 1.65 mg/liter) proceeding from the subsurface to the bottom portion of the water. There is little iron in the lake water (up to 0.2 mg/liter). It is primarily represented by complex organomineralic compounds. In the nearshore part of the lake (see figs. 4 and 5, stations 3, 5, etc.) there are gray and light-gray fine-grained carbonate muds, including grains of quartz, feldspar, well preserved plant remains, and larger fragments of carbonate rock. Mollusk shells belonging to *Planorbudae*, *Limnaea*, and *Viviparus* are fairly abundant among the sediments.

Slightly altered aragonite and calcite enter into the compositions of the carbonates of the shells. The somewhat reduced reflections of a number of peaks and asymmetries of peaks indicate alteration of the aragonite.

Abundances of Fe, Mn, P, C_{org} , CO_2 , and SiO_2 in the calcareous sediments of the nearshore stations are fairly typical (see Table 2). Almost all components with the exceptions of C_{org} and CO_2 , are present in small quantities. Even if these determinations are corrected for carbonate-free material, Fe, Mn, P, and $\text{SiO}_{2(\text{amorph})}$ will be at clarke values or even somewhat below them.

The somewhat elevated content of C_{org} in comparison with that found in nearshore sediments of other lakes is most likely due to presence of relatively coarse plant remains which were resistant to breakdown and which arrived here via an allochthonous pathway, primarily from peat bogs and marshes located near the lake. Sediments located in deeper parts (in the hypolimnion) of Lake Saad'yarv (stations 1, 2, 4) are represented by clayey-limey muds and contain more, probably allochthonous, organic matter. Under the microscope, one can see a reddish-brown lumpy groundmass devoid of structural components and with indistinct remains. The remains are difficult to identify. Such an atypical and, as it were, unnatural in Strakhov's view [25]) co-occurrence of minerals and organic matter may be explainable by the fact that the latter are present here in the form of humates rather than in a free state [3, 18, 21].

Humic acids lose their corrosive properties when bound to Ca; therefore, transported fine-grained carbonate sediments are accumulating here despite elevated C_{org} content at depths.

Use of a combination of methods (staining of thin-polished sections, as well as X-ray, thermal, and chemical analyses) allowed us to establish that the carbonate sediments of Lake Saad'yarv are composed of several minerals, rather than a calcite alone, as many investigators had suggested. Staining of sections of carbonate muds collected from various lake depths directly reveals the presence of different minerals taking on intensive red coloration (calcite), weakly staining crimson (low-magnesian calcite), and resisting staining (high-magnesian calcite and protodolomite).

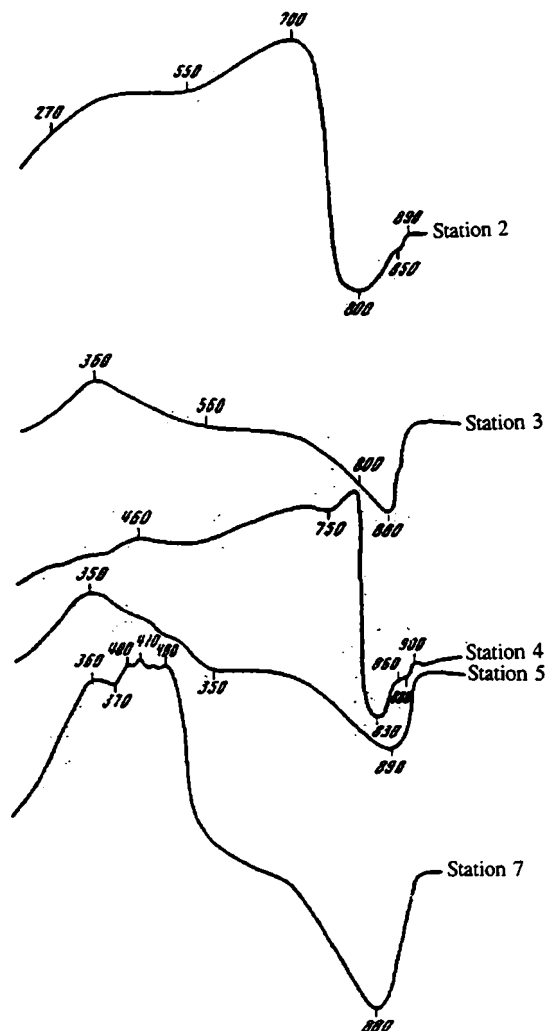


Fig. 6. Heating curves of sediments from Lake Saad"yarv.

Heating curves of carbonate muds collected at shallow water stations 3 and 5 (Fig. 6b, d) are fairly similar. They show a single, clearly expressed exothermal maximum in the region of 350-360°C, corresponding to decomposition of organic matter, and a single endothermal interruption, which may indicate the presence of low-magnesian calcite or fine-grained calcite combined with a relatively large quantity of clay-mineral admixture. The character of the carbonate minerals is revealed more clearly with the aid of X-ray analysis. On the diffractometer curve of a sample of light-gray carbonate mud from Station 3, there is a single principal peak at 3.0297 Å, corresponding to calcite or calcite with a very small isomorphous impurity of Mg (Fig. 7a). This peak is slightly asymmetrical. The branch toward larger angles is somewhat bent. This allows us to assume the presence of carbonates of a more complex composition than calcite at sites where branches of peaks show inflections. Measurements of angles at sites of inflection on the left branch of the curve indicate that, in addition to calcite, there is yet another carbonate mineral present: low-magnesian calcite, showing greatest intensity on the X-ray diffraction pattern at 3.006 Å. Most likely, it is precisely this carbonate which takes on a weakly expressed crimson color when sections are stained with a complexing reagent. The heating curve of the sediment (see Fig. 6c) raised up at Station 4 (from a depth of 10.4 m) differs from the heating curves of sediments collected at shallow water stations 3 and 5 (depths of 1.3 and 1.5 m). A number of endothermal breaks (750, 830, and 870°C), indicating the presence of not one but several carbonate minerals, including the proposed high-magnesian calcite, can be traced in the high-temperature region. On the X-ray diffractometer curve (see Fig. 7b) of sediment from Station 4, it is evident that the principal reflection is located in the vicinity of 2.0297 Å; this corresponds to calcite. In considering the configuration of

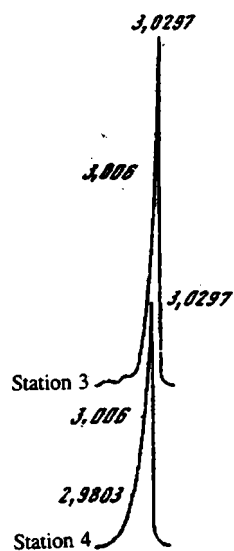


Fig. 7

Fig. 7. Diffractograms of mud from Lake Saad"yarv.

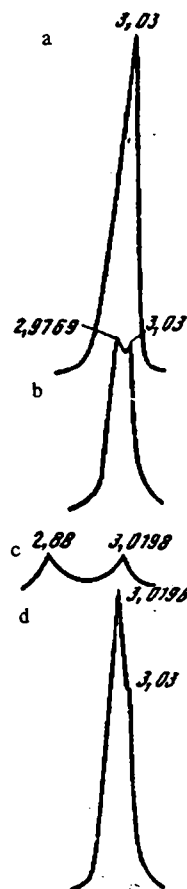


Fig. 8

Fig. 8. Diffractograms of carbonates obtained as a result of experiments: a) tapwater; b) 0.03 N solution of FeSO_4 ; c) 0.04 N solution of MnSO_4 ; d) 0.03 N solution of MgSO_4 .

this peak, note that it is somewhat asymmetrical. Its branch toward the region of high angles has a number of slight inflections, allowing us to assume the presence of other, more magnesian carbonates. Interplanar spacings at points of inflection on the peak's branch equal 3.006 Å (low-magnesian calcite, staining slightly crimson) and 2.9803 Å (relatively high-magnesian carbonate not reacting to the complexing reagent). Consequently, low- and high-magnesian carbonates, in addition to calcite, are present in the sediment from Station 4. Thus, sediments collected from the bottom of Lake Saad"yarv are represented not only by calcite, as most investigators had assumed (investigators who have advocated chemogenic formation of carbonates in lakes of the Baltic region).

Just as in the case of Lake Yantu Siniyarv, we calculated the equilibrium constant K according to Kol'tgof's formula for a number of water samples collected from different stations and depths. In all instances, it turned out that values obtained for K were two and a half orders of magnitude less than Kol'tgof's value of K ; this indicates that direct precipitation of carbonates from the lake water is impossible. Taken as a whole, this suggests a primarily terrigenous origin for carbonate material containing various minerals in sediments of Lake Saad"yarv and other lakes of the Baltic region, where carbonate-containing rocks occur in the catchment areas.

In our opinion, this point of view is entirely correct, since, the carbonate fragments we found among the sand grains in the nearshore areas of Lake Saad"yarv, are, in terms of mineral composition, generally indistinguishable from carbonates found in the center and other areas of this body of water.

Whereas the pathway of carbonate accumulation is, for the most part, clearly terrigenous in the case of Lake Saad"yarv, and, most likely, for many other similar bodies of water in Estonia, Latvia, and Lithuania, the mechanism of

crust formation at bushes located on the bottom of Lake Yantu Siniyarv and, probably, the formation of the muds themselves, remain a mystery to us. In other words, while acknowledging the participation of blue-green algae in these processes, we do not know which types of biogenic precipitation to emphasize as important [17]. To resolve this problem, we carried out the following experiments: algal colonies (primarily *Microcolous* and possibly also *Colothrix*) grown on carbonate crust (from Lake Yantu Siniyarv) in tapwater were transplanted to four 200-ml flasks. We poured the following into the flasks: a) tapwater; b) 0.03 N solution of FeSO_4 in tapwater; c) 0.04 N solution of MnSO_4 in tap water; d) 0.03 N solution of MgSO_4 in tapwater.

Over the course of two weeks, the pH of the water in the flasks fell to 8.9-10; Eh fell to +(460-480) mV. A thin coating of carbonate material was visible on the bottoms and the sides of the flasks. An X-ray diffractometric analysis of the carbonates formed in the flasks is presented in Fig. 8. When blue-green algae were grown in tap water, the precipitate proved to be calcite (see Fig. 8a). When FeSO_4 was added to the water, two carbonates formed: calcite and slightly ferruginous calcite (see Fig. 8b); this was confirmed when the powder was stained with a complexing reagent. When MnSO_4 was added to the water, two minerals also appeared in the precipitate, low- and high-manganese calcite (see Fig. 8c); finally, addition of a MgSO_4 solution to the original tapwater resulted in appearance of two minerals (see Fig. 8d): low-magnesian calcite (primary) and calcite (secondary). In our opinion, this experiment satisfactorily proves that blue-green algae play a direct role in precipitation of carbonates by alkalizing the medium (increasing the pH). Thus, it can be stated that precipitation of carbonates in Lake Yantu Siniyarv most likely takes place via a biochemical pathway involving the participation of blue-green algae, rather than via a purely chemical pathway.

Let us add that our investigation of lakes of Lithuania, Latvia, and the Leningrad Oblast, the results of which were not included in this article, and an analysis of published data concerning carbonate formation in these regions also confirms the important role of terrigenous accumulation of carbonates and the secondary role of biochemogenic origin.

* * *

Our investigation of features of carbonate formation in lakes in the north of the Russian platform shows that overland accumulation of carbonates (which is usually represented by other minerals besides calcite), rather than chemogenic deposition, predominates in basins with hard water situated among carbonate rocks in catchment areas. These carbonates frequently coexist in the hypolimnion along with allochthonous, distinctly unreactive organics.

A generally similar point of view has been put forward by Baker-Blocker and Collender [31], who investigated one of the southern inlets of Lake Michigan.

An analysis of carbonate accumulation in the lakes of Minnesota [32, 33] also suggest that the carbonates found in these bodies of water are of overland origin. Biochemogenic accumulation as well as overland accumulation of carbonates takes place in lakes of the northern part of the Russian platform. On the one hand, there is the removal of CaCO_3 from water by organisms for construction of shell skeletons, noted earlier by many investigators and, on the other hand, the appearance of carbonates as a result of the life activity of blue-green algae. In general, these processes are weakly developed and manifest to the full extent only in favorable situations (during an absence or sharp drop in supplies of terrigenous material, reduced accumulation of silica, a minor role of organic in sedimentation, etc.).

Evidence indicating incorporation of manganese into the compositions of biogenically formed carbonates is also important. Whereas manganese utilization by many carbonate-precipitating organisms (foraminifers, algae, etc.) in marine basins has been acknowledged, the presence of manganese within the composition of neogenic carbonates under fresh water conditions, in our opinion, has not been considered previously.

In summary, it can be said that the terrigenous pathway is the main pathway involved in accumulation of carbonates in bodies of water located the territory of the northwestern platform, where carbonate rocks occur in catchment basins and the waters of rivers and lakes are essentially hard.

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LOCALIZATION PATTERNS OF COPPER MINERALIZATION IN THE UDOKAN DEPOSIT

I. F. Gablina and A. I. Mikhaylova

UDC 553.43(571.5)

In the article, we discuss location patterns of copper mineralization in the Udokan deposit. The main patterns are restriction of mineralization to a metamorphosed molassoid red-bed formation infilling a large intermontaine trough and the outlining of the mineralization by lenses of grey pyritiferous rocks serving as a geochemical barrier for ore formation. We show that the primary ores formed during pre- and syn-metamorphic stages were subjected to thorough supergene reworking, leading to enrichment of the ores with secondary sulfides.

Udokan is one of the oldest deposits of cupriferous sandstones and slates. Investigation of the conditions of its formation is therefore important not only from the point of view of the theory of stratiform ore formation but also in order to reconstruct features of the geological-geochemical environment in the Precambrian. Numerous reports [7, 8, 10-12, 15, etc.] have discussed the geological structure, composition of ore-enclosing rocks, and mineral features associated with the mineralization of this deposit. At the same time, a number of unresolved or controversial questions continue to hinder reconstruction of a complete picture of its formation. Above all, these include questions concerning the formation classification of the ore-enclosing deposits and their depositional environment, concerning primary ore-controlling factors, concerning the role of superimposed processes in the formation of the deposit, and concerning the mineral zonation evident in the mineralization and the circum-ore alterations.

In this article, we have undertaken a comprehensive analysis of the localization patterns of mineralization in the Udokan deposit, based upon a compilation of data from previous investigations and original data obtained by the authors.

GENERAL CHARACTERIZATION OF THE UDOKAN DEPOSIT

The Udokan deposit is restricted to the early Proterozoic Udokan series infilling the large Kodar-Udokan trough. The geotectonic setting of the latter has been variously interpreted: as a proto-platform structure (by A. M. Leytes) or as a miogeosyncline (by L. I. Salop), but the majority of investigators classify the copper-bearing deposits of the Udokan series deposited within the trough as orogenic molassoid formations.

The Udokan series is a 10-12-km sequence of uniform grey rocks, predominantly of aleuritic-arenaceous composition. Rudaceous and even carbonate and argillaceous rocks play a predominant role. The basement of the trough is composed of thoroughly metamorphosed crystalline rocks of the Chara series of the Archean and greenschist-altered sedimentary-volcanogenic deposits of the trough complex. The latter are formations of geosutural troughs, i.e., rifts and riftogenic belts [6]. In the opinion of the majority of the investigators, the Kodar-Udokan trough was, during the time of deposition, separated by a transverse high into two subzones of almost identical area (Fig. 1): the Kodar (in the northwest) and the Udokan (in the southeast). The synsedimentary high became exposed as the thickness of the Udokan series above it decreased [18].

The following groups can be distinguished within the section of series in the Udokan subzone (bottom to top): the Ikabiysk, Ayansk, Inyrsk, Chitkandin, Aleksandrovsk, Butun, Sakukan, and Naminga. The Udokan series reaches maximum thickness (10-12 km) in the centers of the Kodar and Udokan subzones. Along the periphery and in the vicinity of the buried high, the thickness of the Udokan series decreases 2-3 fold (see Fig. 1).

Rocks of the Udokan series are mounted into large-scale brachy-folds complicated by later deformations and the development of faults. Injections of intrusive rock masses and dike bodies have taken place along the faults.

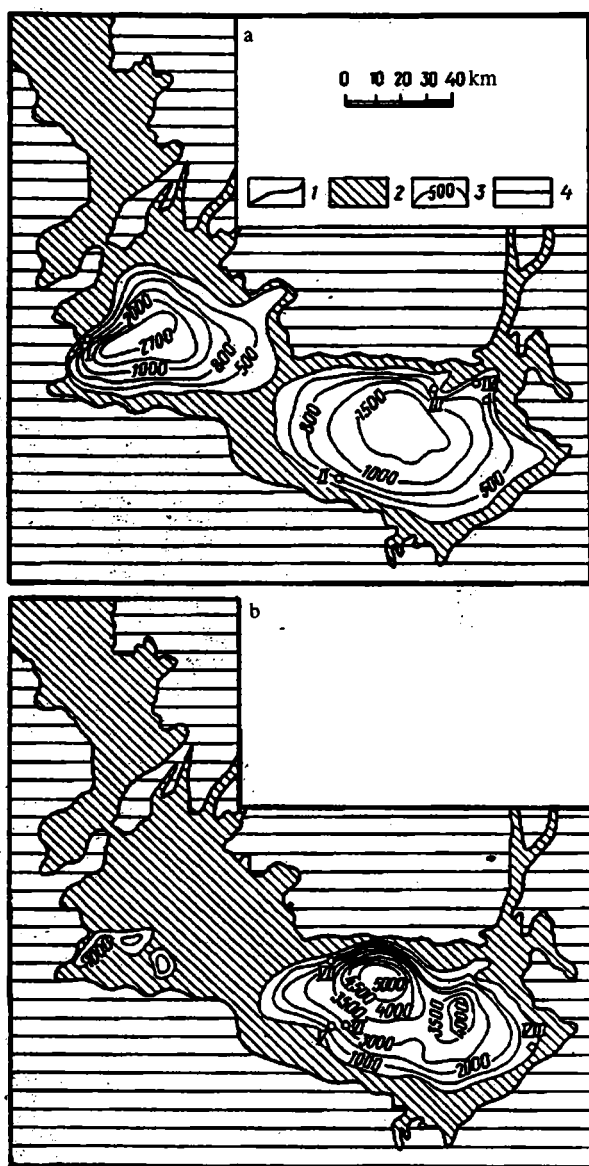


Fig. 1. Schematic thickness map of the Udokan (a) and Sakukan (b) groups in the Kodar-Udokan trough [17] and the locations of copper shows within the trough: 1) outline of the trough in the present-day eroded section; 2) areas within the trough where deposits of the group are absent; 3) isopachs, m; 4) territory outside the trough; I-VIII) deposits in oreshows (I: Syud'ban, II: Pravyy Ingamakit, III: Krasnoye, IV: Khani-Sakukan, V: Udokan, VI: Saku, VII: Unkur, VIII: Burpala).

The rocks are metamorphosed to the greenschist facies (quartz-chlorite-muscovite-epidote-biotite paragenesis) and, in part, amphibolite facies (the appearance of actinolite, hornblende, and scapolite). The rank of metamorphism increases downward along the section and toward contacts with granite gneisses (along the periphery of the trough). Investigators link the metamorphism of the Udokan deposits to the stage of early Proterozoic granitization of the Archean basement and, in part, to contact interactions with later intrusions penetrating the Udokan series.

Platform formations of the Late Proterozoic, Cambrian, and Ordovician lie upon rocks of the Udokan series in the southeastern part of the trough. Jurassic and Cretaceous lacustrine-terrestrial coal-bearing deposits formed in separate, narrow grabens overlies these formations with an erosional break and basal conglomerates [17].

Copper mineralization in the Kodar-Udokan trough has been discovered at several stratigraphic levels of the Udokan series: the Chitkandin (the Pravyy Ingamakit, Krasnoye, and other oreshows; the Aleksandrovsk, Lower Sakukan (the Burpala, Unkur, etc.),* and the Upper Sakukan (Udokan, Saku). Their spatial distributions are controlled by the trough borders and the slope of the transverse high (see Fig. 1).

The Udokan deposit is restricted to the upper copper-bearing level of the Udokan series, the Upper Sakukan subgroup. It is located within the Naminga syncline, situated in the southwestern part of the Kodar-Udokan trough. The Naminga syncline takes the form of an ellipse elongated in a northwesterly direction (Fig. 2). In the southwest, the rocks of the Udokan series compose a syncline cut by gabbroids of the Chineya massif; in the northeast, they compose the Kemen granite massif belonging to the Kodar complex. The middle and upper subgroups of the Sakukan group play a role in the structure of the deposit; they compose the limbs of the syncline, with the Naminga group extending into its core. The rocks dip toward the center of the structure at angles ranging from 12-20° to 35-42°. Only in the central part of southern limb is the bedding of the rock overturned, with beds dipping toward the southwest at angles ranging from 30° to 80°.

Dike formations of various ages and compositions are abundant within the vicinity of the deposit. The largest is the Glavnaya gabbro-diabase dike of Early Proterozoic age (70-180 m thick), criss-crossing the entire deposit from the northeast to the southwest and steeply dipping toward the northwest at angles of 75-83°. The gabbro-diabase dike and the enclosing sedimentary rocks are crosscut by a thinner (10 m) lamprophyre dike tentatively assigned an Mesozoic stage of activation. The lamprophyre dike cuts across the entire extent of the deposit in a meridional direction and dips toward the west at angles of 55-68°.

Faults can be subdivided into three main systems: 1) older faults of northeasterly trend; 2) faults of northwesterly and meridional trends, tentatively assigned to the Mesozoic; 3) faults of sublatitudinal trend and problematic age, with the majority renewed during the Cenozoic.

The absence of Late Ordovician sediments in the region suggests that most of the territory served as a area of crushing, starting with the Silurian. The Paleozoic—Mesozoic transition, marked the beginning of the stage of tectonic platform activation, which resulted in a dissected topography continuing into the Jurassic and Cretaceous [17]. The territory was peneplained at the end of the Mesozoic. The zone of weathering from this time was preserved in places outside the deposit under sheets of Pliocene plateau basalts. Remnants of ancient (Pre-Pliocene) surfaces of alignment with evidence of ancient valleys reworked by glaciers are widespread in the vicinity of the deposit [8, 12].

From the second half of the Pliocene, a new period of uplift began, accompanied by intensification of erosional activity and formation of erosional topography, which, during the Middle Quaternary, was reworked by glaciers of the second stage of glaciation.

The Late Quaternary was the occasion of the most recent movements, resulting in the formation of the present-day alpine topography.

The copper ores of the Udokan deposit include a sulfide dissemination which is oxidized to a substantial degree. The main ore-forming minerals are sulfides of the chalcocite series (djarleite and digenite), bornite, chalcopyrite, and covellite. Magnetite, hematite, and pyrite have the greatest distributions in the ores. Chalcosine, anilite, molybdenite, stromeyerite, wittichenite, rutile, sphalerite, marcosite, pyrrhotite, etc. are occasionally found. Minerals of the oxidation zone, especially oxides and hydrous oxides (goethite, psilomelane, occasionally tenorite and lepidocrocite), sulfates (brochantite, chalcantite, antlerite, and jarosite), and carbonates (malachite and azurite) are widespread within the deposit. The sulfide minerals form disseminations and bedded accumulations of various thicknesses, primarily along the borders of rocks of a different granulometric composition, but also around fragments of argillites and aleurites in sandstones. The sulfide minerals emphasize the original sedimentary textures of the rock. At the same time, most of the sulfide formations have corrosion-related forms of segregation superimposed upon the country rock. The veinlet type of mineralization is less common. Veinlets range from 2 mm to 1.5 cm in thickness; thick ore veins are less common; they mainly occur near dike bodies. They contain the same primary copper ores that are found in the disseminated ores.

Oxidized minerals inherit textural features of sulfide ores and also form incrustations along cracks and thin veins in crush zones as well as on the walls of mines.

*Recently, several investigators have assigned these ore shows to the Chitkandin level [1, etc.].

Over most of the deposit, ore beds take the form of lenticular sheeted bodies with fairly persistent thickness and content. On the north rim of the syncline, near near exposure at the surface, the structure of the ore bodies are complicated as a result of thick bulges, offshoots, abrupt wedge-outs, and pinch-outs. Bulges of ore bodies and an increase in the overall thickness of the ore horizon are evident near the Glavnyy dike.

ORE-ENCLOSING DEPOSITS OF THE UDOKAN DEPOSIT

The middle subgroup of the Sakukan group is 1500-1600-m thick in the vicinity of the deposit and is characterized by a uniform sandstone composition, cross-laminated rock texture, and an abundance of layers rich in magnetite.

The upper subgroup is an interbedding of intermediate- and fine-grained, grey, occasionally pink sandstones, often containing inclusions of angular fragments and debris of siltstones, mudstones, and dark grey fine-grained quartzitelike sandstones. The sandstones are cross-grained, with interlayers of native alluvial sand showing a predominance hematite, martite, magnetite, and sphene developed from fragments of titanomagnetite and chromite [5]. Fragments of garnets, zircon, tourmaline, disthene, pyroxenes, and other minerals of the heavy fraction are present in substantially smaller quantities. In the middle of the subgroup, there is a lens of interstratified sandstones with siltstones and mudstones. Copper mineralization is confined to this lens (the ore member). It exhibits a clearly rhythmical structure expressed in an alternation (from below upward) of intermediate-grained calcareous sandstones with siltstone and mudstone clasts, fine-grained dark grey quartzite-like sandstones and interlayers of mudstone and siltstone several centimeters to 5 m thick. At base of the rhythms, one typically finds erosion surfaces and lenses of intraformational conglomerate-breccias. Rhythms rarely exceed 5-10 m in thickness. The lens showing interstratification is not persistent in thickness or strike. It pinches out in the northwest and southeast directions (in areas of centroclinal closure of the syncline) as a consequence of disappearance of mudstone and siltstones interbeds from the section; these interbeds are replaced by magnetite-bearing intermediate- and fine-grained sandstones with inclusions of siltstone clasts. These decrease in number toward the southern border of the syncline, where lenses of intraformational conglomerate-breccias, sometimes reaching 10 m in thickness and extending for more than 500 m, become abundant.

The interbedded lens shows a maximum thickness (~400 m) toward the northern border of the Naminga syncline, where it is located 200-300 m below the foot of the Naminga group. In all directions from this site (west, south, east), the ore-enclosing member decreases in thickness and is offset diagonally upward along the section. Thus, it is reduced to a thickness of 250-100 m toward the southern border of the syncline, where it is located 120-150 m from the foot of the Namingina group.

The maximum thickness of the upper Sakukan subgroup, 600-650 m, is found at the northern border of the syncline. The thickness of this subgroup decreases to 200-300 m toward the southwest and east. Substantial subsidence of deposits of the subgroup is occurring west of the Naminga syncline. The copper-bearing upper Sakukan subgroup of the Naminga group is predominantly of an aleurolitic composition and is 800-1000 m in thickness.

Regional metamorphism of the ore-enclosing deposits corresponds to lower ranks of the greenschist facies. The rocks have preserved primary-sedimentary structures and textures. Metamorphism is expressed in the substantial recrystallization of the cement of the sandstones and formation of a mineral paragenesis typical of this facies (quartz, feldspar, muscovite, chlorite, epidote, and biotite).

The contact metamorphism associated with the Kemen granitoid massif is primarily expressed in epidotization, found over the entire territory of the deposit and occurring most concentrated form within a 300-400-m aureole from the contact. At the contact with the Kemen granitoid massif, sandstones of the Sakukan group have become metamorphosed hornfels (for several meters). Exocontact alterations associated with a gabbro-diorite dike occur within 15-20 m from the contact and are expressed in the occurrence of neogenic amphibole and feldspar, as well as in an increase in the quantity of epidote.

THE DEPOSITIONAL ENVIRONMENT AND SOURCES OF DRIFT FOR THE ORE-ENCLOSING DEPOSITS

A hypothesis put forward by Bogdanov et al. [10] was considered plausible until recently. Bogdanov suggested that the ore-enclosing deposits of Udokan formed under conditions of a delta with large-scale flow carrying water and

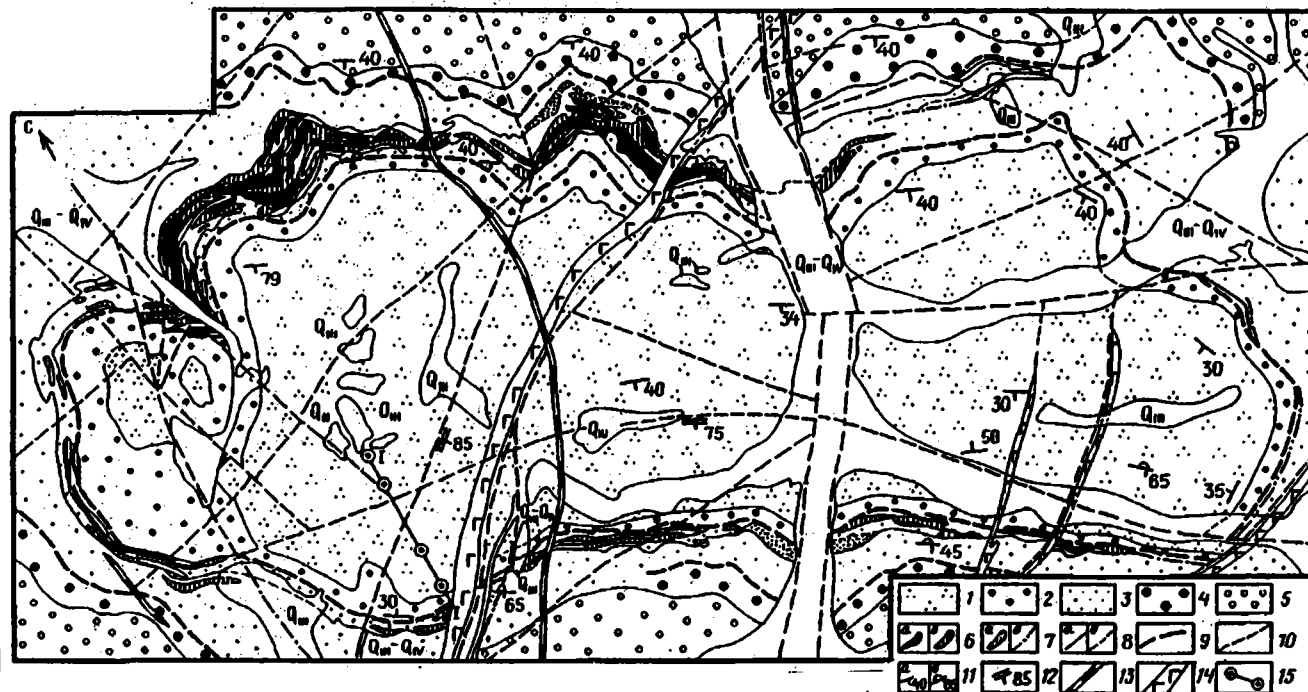


Fig. 2. Geological map of the Udokan deposit (according to R. N. Volodina, V. G. Evsyukova, and V. S. Chechetkina). 1-5) Deposits of the Udokan series of the Lower Proterozoic (1: Naminga group, 2-4: upper subgroup of the Sakukan group of the supra-ore, ore, and sub-ore members, respectively, 5: middle subgroup of the Sakukan group); 6) copper ores (a: rich, b: lean); 7) sandstones with lean mineralization (a: relatively thick, b: thin beds); 8) geological boundaries (a: established, b: tentative); 9) boundaries of ore member; 10) faults; 11) attitude of rock (a: normal, b: overturned); 12) bed positions of faults; 13) dikes of lamprophyres and quartz porphyries, tentatively dated as Mesozoic; 14) dikes of gabbro-diabase of the Chineya intrusive complex of Upper Proterozoic age; 15) line of profile with boreholes; Q_{III-IV}) recent alluvial and Upper Quaternary aqueo-glacial deposits; Q_{III}) Upper Quaternary glacial deposits.

sedimentary material from a land area (the Chara ridge) toward a hypothetical marine basin to the south. However, later research has turned up evidence against the idea of a single direction of washdown. These facts include:

1. The distribution of thicknesses of deposits of the Upper Sakukan subgroup (see Fig. 1) suggests that the region of sedimentation was a closed basin during the deposition of the ore-enclosing stratum. Its outlines were similar to the present-day boundaries of the Udokan trough; a zone of maximum downwarping was located in the northern part [16, 17]. A similar pattern can be traced in the distribution of thicknesses of the ore-bearing horizon itself, which takes the form of a lens which bulges toward the northern border of the synclinal structure and gradually pinches out in the southern, eastern, and western directions.

2. The distribution of the lithofacies of the ore-enclosing deposits. Very fine-grained silty-sandy facies of the Upper Sakukan subgroup occur in the northern, most downwarped, part of the basin, with the facies grading into more coarsely fragmental, sandy formations in all directions from that area. In the western part of the Kodar-Udokan zone, there are even gravellites and conglomerates [15, etc.], suggesting proximity to the source area.

The data presented suggest the formation of the ore-enclosing deposits in a closed synsedimentary basin of asymmetrical shape, with the center of downwarping shifted toward the north, where the fine-grained sediments were deposited. Washdown of material may have taken place from all sides. The shoreline of the basin closest to the Udokan deposit was probably located to the south and west.

The petrographic composition of the ore-enclosing fragmental rocks suggest formation of the rocks was substantially a result of breakdown of volcanogenic complexes. Sandstones of the Sakukan group contain 15-30% rock fragments, including fragments of acidic effusives, jasper, and granophyres. Pebbles of puddingstones and gravellites from the middle subgroup of the Sakukan group show a similar composition [2, 9]. Individual clasts of pyroxene and hornblende as well as recently found relicts of chromite partially replaced by magnetite and restricted to interlayers of native alluvial sand are also preserved, together with clasts of acidic volcanic and igneous rocks in sandstones of the Sakukan suite [5]. The minerals mentioned suggest the presence of ultrabasic magmatic rocks in the source area, with most of these rocks not being preserved in the metasedimentary deposits of the Sakukan suite because of their instability under catagenic conditions. In particular, basic and intermediate plagioclases were subjected to intra-layer dissolution and albitization during catagenesis, whereas titanomagnetite was decomposed into hematite and sphene. The decomposition products of the minerals mentioned (albite, hematite, and sphene) are widely occurring in sandstones of the Sakukan group.

In the opinion of Fedorovskiy [18], the sources of clastic material may have included rocks of the main greenstone belt, located in the Central Vitim highland, west of the Kodar-Udokan trough, as well as volcanoplutonic and ferruginous-siliceous formations of the trough complex located directly within the trough borders. Grabbroids and ultrabasites, potential sources of titanomagnetite and chromite, are widely occurring in these and other formations. In the opinion of Burmistrov [2], meta-effusives of the trough complex differ from fragments of effusives contained in the Sakukan group in exhibiting a higher rank of metamorphism, on the basis of which, the author has concluded that they are a less likely source of washdown. Taken as a whole, the configuration of the Udokan basin during Middle and Upper Sakukan times (see Fig. 1) suggest that clastic material for the ore-enclosing deposits may have entered from the entire rim of the basin. The increase in roughness of clastic material in the western part of the Kodar-Udokan trough suggest that the source area was closest in this area.

As follows from the description presented above, the copper mineralization of the Udokan series was restricted to a large intermontaine basin, the Kodar-Udokan trough. The spatial distribution of copper shows within the latter is controlled by the trough borders (see Fig. 1). The largest Udokan deposit probably abuts the southern slope of a buried high separating the Kodar-Udokan trough into two parts.

FORMATIONAL CLASSIFICATION OF THE ORE-ENCLOSING DEPOSITS

There are various points of view concerning the formational classification of the ore-enclosing deposits of Udokan. Many investigators (A. I. Leytes, M. E. Berdichevskaya, N. V. Sochava, Yu. V. Bogdanov, V. P. Feoktistov, etc.) group them into an orogenic redbed formation which formed in a subaerial environment and which lost its original appearance during metamorphism. Grintal' [7] and Salikhov [14] suggest that the grey color of the ore-enclosing stratum is the original color and was a result of the reducing environment of sedimentogenesis. Gablina [4], on the basis of mineralogical-

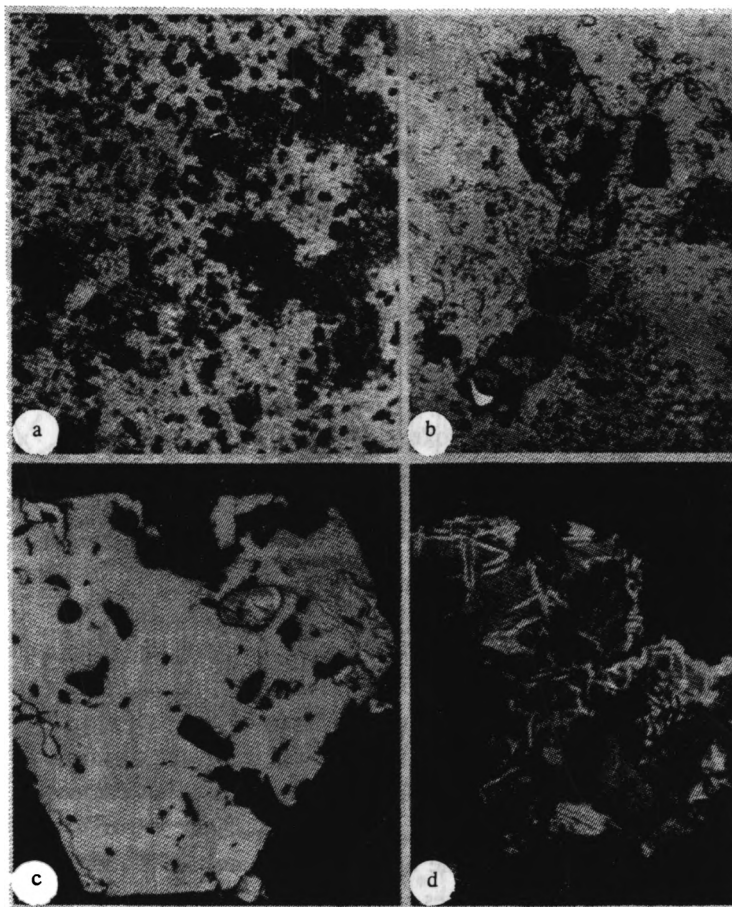


Fig. 3. Microphotographs of thin sections (a, b, without analyzer) and polished sections (c, d): a) fine dissemination of pyrite on epidotized argillite. In areas of epidotization (grey), pyrite is replaced by epidote, magn. 200; b) replacement of pyrite-chalcopyrite paragenesis by epidote in ore-bearing sandstone, magn. 200; c) inclusion of bornite crystal (light-grey), partially replaced by covellite (dark-grey) in a metacrystal of magnetite (grey), magnetite with edges replaced by late hematite (white), magn. 200; d) replacement of bornite (grey) by digenite (light-grey) and covellite (dark-grey), accompanied by segregation of blades of synantetic pyrite (white), magn. 200.

geochemical criteria (the ratio $\text{FeO}/\text{Fe}_2\text{O}_3$ and the presence of indicator minerals) in the Udokan series, distinguishes between complexes of originally grey and originally red rocks alternating in the section. The upper part of the section of the Udokan series, including the Udokan deposit (the Middle and Upper Sakukan subgroups and the Naminga group) belong to originally red-colored deposits. The indicator minerals of an oxidizing environment for their formation are authigenic hematite and magnetite disseminated in the rock. The hematite and magnetite formed during catagenesis and metamorphism as a result of the presence of primary finely dispersed iron oxides (the pigmenting substance) according to the pattern: ferrihydrite-hydrohematite(?)—hematite-magnetite. Loss of the original red coloration was caused by coarsening of hematite crystals as a result of collective recrystallization, which began during catagenesis [4]. Conversion of hematite to magnetite is possible at temperatures around 400°C ; this corresponds to temperature conditions of the greenschist facies of metamorphism. Reducing agents, possibly H_2 , CO , and sulfides, had a stimulating influence on this conversion. Our investigation showed that the maximum contents of metamorphogenic magnetite at Udokan are typical of calcareous and ore-bearing sandstones, i.e., rocks containing reducing agents.

Pyrite is an indicator of a reducing environment in which initially grey-colored rocks formed.

FACTORS ASSOCIATED WITH PRIMARY CONTROL OF MINERALIZATION

As indicated by our investigations, the interbedded member to which copper mineralization is confined at Udokan (the ore member) contains from 1 to 3 lenses of pyritiferous rock. Pyrite forms a fine dissemination in interlayers of very fine-grained rock: mudstone and siltstone (Fig. 3a); this is reason to classify it as a syndiagenetic formation. Pyrite often enriches individual, thin interlayers in sandstones and is reminiscent of recrystallized accumulations of clastic grains. The presence of diagenetic and clastogenic pyrite suggests a reducing environment for the sedimentation and diagenesis of these rocks. Still later (catagenic and metamorphogenic) formations of pyrite in the form of metacrystals join syngenetic segregations of pyrite in sandstones of the ore and subore members. Zones of occurrence of catagenic pyrite form aureoles around syngenetic pyrite-bearing lenses and show blurred boundaries which do not coincide with lithological boundaries.

Lenses of rock with pyrite are offset en echelon upward along the section from the north, where the ore member is maximally thick. In plan, the zone of pyrite-bearing rock takes the form of a broad belt extending in a meridional direction and pinching out toward the western and eastern flanks of the deposit. Copper mineralization pinches out in the same direction (to the west and east). From the south and north, it is cut by zones of denudation on the limbs of the Naminga syncline. Copper mineralization in the form of chalcopyrite, bornite, and chalcocite rims pyritic lenses in plan and section (Fig. 4a). The presence of two or more lenses with pyrite results in the formation of multi-stage copper mineralization. In the process, pyrite is replaced by copper-containing sulfides in the sequence: chalcopyrite, bornite ($\pm$ sulfides of the chalcocite series).

Thus, the copper mineralization of the Udokan deposit is controlled by interlayers of originally light-grey pyrite-containing rocks within a sequence of red terrigenous deposits which lost their original color during metamorphism. Apparently, lenses of pyritiferous rock serves as reducing barriers during oxidation of sulfide ores. Both hydrogen sulfide, present in the originally grey interlayers, and diagenetic segregations of pyrite could have acted as oxidizers of copper. It is known that iron (II) sulfides can serve as reducers for divalent copper in solution. This takes place under conditions when the pH of a solution containing copper exceeds the range of stability for pyrite. In addition, displacement of iron by copper may be due to the fact that copper has greater affinity for sulfur than iron has. Widespread occurrence of reducing reactions of copper with pyrite in nature is suggested by the fact that structures created by replacement of pyrite by copper sulfides are ubiquitous in ores of copper sandstones and slates.

PARAGENETIC ASSOCIATIONS OF ORES MINERALS

Pyrite is the earliest sulfide mineral. All of the above-mentioned morphogenetic types of pyrite (clastic accumulations, diagenetic fine disseminations, and metacrystals) are replaced by copper sulfides as well as magnetite, epidote, and other metamorphic minerals.

Among the copper sulfide minerals, we have identified several non-coeval paragenetic associations which can be assigned to different stages of lithogenesis and metamorphism of the enclosing rock and to the supergene process. They can all be clearly subdivided into two main groups: pre- (or syn-) metamorphic and post-metamorphic. The classification criterion is their inter-relationship with metamorphic minerals of the enclosing rock (biotite, epidote, magnetite, etc.). Sulfides of the early paragenesis are represented by chalcopyrite, stoichiometric bornite, digenite, and chalcocite with an admixture of tetragonal forms of Cu_2S . They are replaced by metamorphic minerals of the enclosing rock (see Fig. 3b), are contained within them in the form of inclusions (see Fig. 3c), or are associated with them, forming sulfide-epidote or sulfide-biotite nests.

The later paragenesis includes djarleite, low temperature digenite, anilite, spionkopite (sic), yarrowite, and covellite. Copper sulfides of the later paragenesis have corrosion-related forms of segregation and are replaced by progressively more durable sulfide minerals and authigenic formations of the country rock, including metamorphic formations. They are associated with late hematite and develop together with it from magnetite. They replace bornite very actively; this is accompanied by segregation of synantetic chalcopyrite, forming blades and lattices of the type associated with breakdown of a solid solution. These blades and lattices are located along the replacement front (see Fig. 3d).

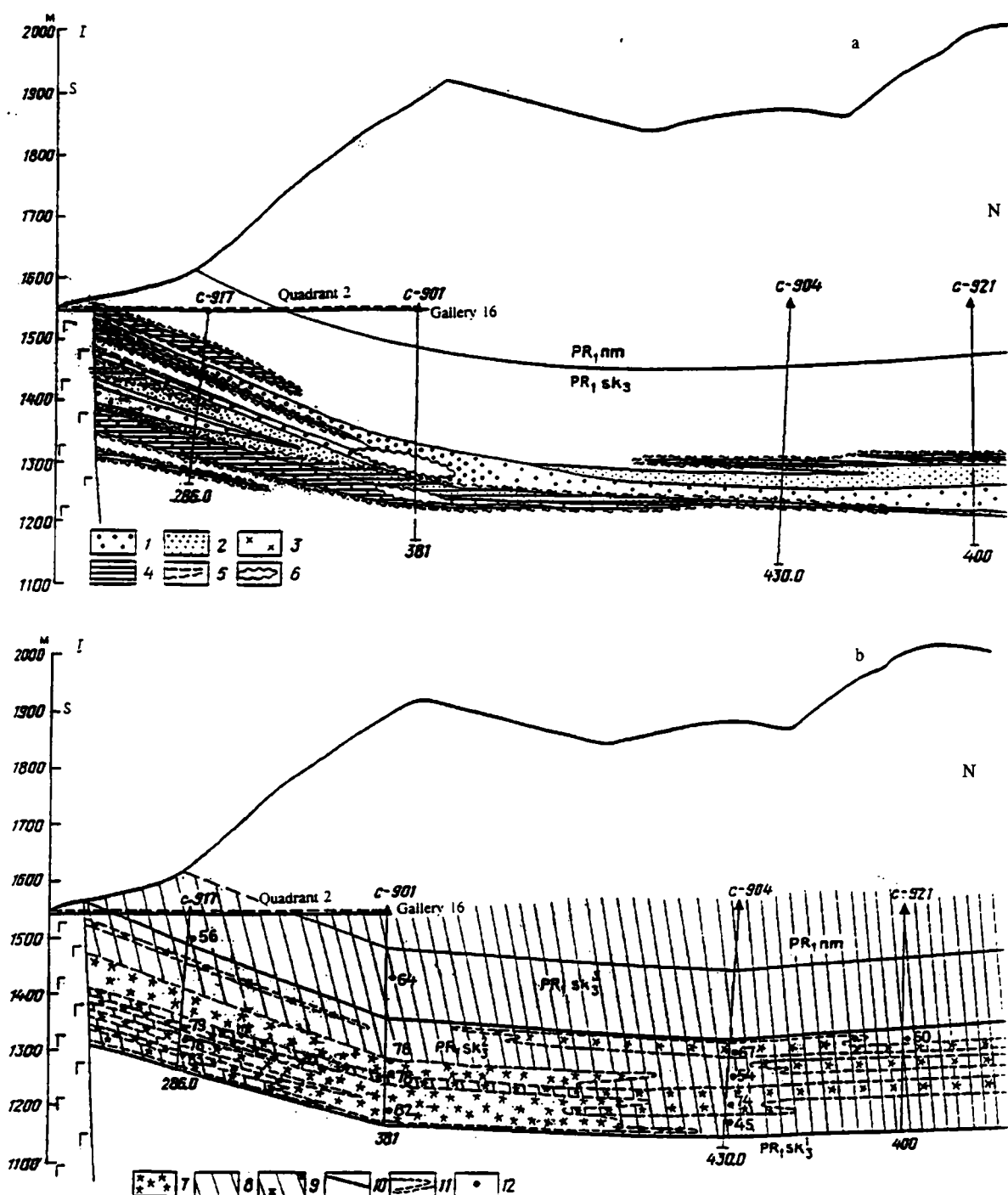


Fig. 4. Scheme showing mineral zoning of mineralization (a) and circumore alterations (b) of the southern flank of the Udokan deposit (after Snezhnyy): 1-3) mineral zones of primary sulfide ores (1: pyrite, 2: pyrite-chalcocopyrite; 3: bornite); 4) secondary sulfides (minerals of the chalcocite series, covellite); 5) outline of distribution of secondary sulfides; 6) zones of leaching; 7) silicified rock; 8) alteration of calcitized and carbonate-free sandstones; 9) same, with sparse interlayers of silicified rock; 10) boundaries of ore member; 11) outlines of zones of secondary alteration of rock; 12) content of SiO_2 in rocks, %.

We have assigned sulfides of the early paragenesis to the stages of catagenesis and metamorphism [3]. Their stability parameters correspond to pT-conditions associated with lower facies of metamorphism. The sulfides of the early paragenesis form a symmetrical mineral zoning expressed in an replacement of pyrite ores with chalcopyrite ones, followed by a replacement with bornite ores, proceeding from the inner parts of ore bodies to their roofs and feet (see Fig. 4a).

The latter paragenesis is widely occurring in the Udokan deposit. Sulfides of the later paragenesis form bulges of ore bodies near their outcrops at the surface and overlap various zones of the early paragenesis (see Fig. 4a), thus disrupting the primary zoning. Nonequilibrium associations (djjarleite + chalcopyrite, digenite + chalcopyrite, etc.); these associations suggest an abrupt change in the physical-chemical environment of ore formation and a lack of subsequent high-temperature interactions. In terms of stability parameters, the sulfides of the late paragenesis and the nonequilibrium associations created correspond to conditions associated with a supergene process and can be classified as formations of a zone of secondary sulfide enrichment.

SUPERGENE ALTERATIONS OF ORES

Thus, our paragenetic analysis and the features of mineral zoning of ore deposits support the controversial suggestion [10] by A. Yurgenson, F. P. Krendelev and others that Udokan has been a site of widespread processes of secondary sulfide enrichment due to prolonged supergene reworking of primary sulfide ores possibly already brought to the surface during the Paleozoic. Various factors have influenced the distribution of minerals in the zone of secondary sulfide enrichment: recent and ancient terrain forms, composition, layering, and attitude of the enclosing rocks, as well as the primary sulfide zoning and morphology of the ore bodies. A distinct secondary vertical zoning is therefore lacking in the deposit.

The action of surficial reworking of primary ores was also expressed in intensive oxidation of sulfides. Minerals of the oxidation zone are found at all hypsometric levels of the deposit, without exception. According to data of the Udokan Mining Service, the average extent of oxidation of ore at the surface is $\approx 70\%$; at the -260-m level, it is 31% , and at the -670-m level it is $\approx 10\%$.

Processes of oxidation and leaching of sulfide ores are maximally developed in sectors of the deposit with very smooth terrain forms belonging to relicts of an ancient surface of alignment.

Jointing of the enclosing rocks played an important role in the oxidation of sulfide ores. Boundaries of distribution of oxidized ores, generally conforming to the outline of the present-day surface, form deep pockets in fault zones. Often, they are accompanied by intensive leaching of ore minerals.

CIRCUM-ORE ALTERATIONS OF THE ENCLOSING ROCK

Against a background of substantial regional and contact alterations of ore-enclosing deposits in the vicinity of the ore deposit, one can detect local circumore alterations: carbonatization, silicification, etc. The most clearly manifest is silicification, as demonstrated in reports by Yurgenson [19] and Petrovskiy [13, etc.]. We carried out detailed investigations of the distribution of silicified rock and discovered that the zone of silicification is controlled by the rhythmically layered ore member and, for practical purposes, does not extend beyond it (see Fig. 4b).

Within the ore member, silicification is layered in character and most commonly is restricted to interlayers of fine- and very fine-grained sandstones located in nonore parts of the rhythms. Such interlayers vary in thickness from 0.1 to 2 m.

The content of SiO_2 in the silicification zones equals 74-83%; the content of SiO_2 outside the zones does not exceed 45-64%.

Silicification is spatially confined to pyritic, chalcopyritic, and bornite zones of copper sulfide ore, reaching maximal occurrence in sectors of the deposit with thick bornite mineralization. Intensive silicification persists in zones of leaching of the mineralization (see Fig. 4).

Silicified sandstones are light green in color, exhibit dense, massive texture, and have a siliceous appearance. In terms of composition, the sandstones are most often feldspar-quartz rocks consisting of clastic grains of quartz, plagioclase, and microcline showing traces of intra-layer solution and regeneration, along with sericitic, quartz-sericitic or quartz cement with flakes of muscovite and, occasionally, biotite. The cement, interstitial and filming-interstitial, comprises 10-

40% of the rock and is unevenly distributed. It probably formed during recrystallization of a primary clayey or silty-clayey mass.

Silicification manifests in the development of fine- and very fine-grained mosaic aggregates of quartz with individual, larger grains. Late quartz is characterized by an absence of exogenic inclusions and rectilinear outlines of grains. Quartz aggregates replace the quartz-sericitic cement of the rock, often capturing clastic grains. In pyritized varieties of silicified sandstones, epigenetic pyrite (or chalcopyrite) typically occurs in close association with neogenic fine- and intermediate-grained quartz and muscovite. In bornite-ore zones, neogenic quartz is often associated with bornite-epidote nests. Near faults, silicification is manifest in the development of numerous thin quartz veinlets.

The degree of silicification of the rock definitely depends upon the granulometric composition. Silicification is confined to areas of relatively coarse-grained deposits. Coarsening of detrital sandstone material and gradual pinching out of the interbedded member (containing interlayers of siltstones and mudstones) in the eastern and western flanks of the deposit is accompanied by decreasing intensity of silicification and predominance of calcitization of processes in the rocks.

The paragenesis of late quartz with metamorphogenic generations of sulfides (pyrite, chalcopyrite, and bornite) and metamorphic silicates (muscovite and epidote) suggest a metamorphic origin for the silicification.*

* * *

The principal location patterns of mineralization in the Udokan deposit are the following:

1. Confinement of the deposit to the slope of a synsedimentary high in a large-scale intermontaine trough.
2. Association with a molassoid red-bed formation deposited as a result of breakdown of metamorphic and volcanic (acidic and basic) complexes containing ultrabasics and deprived of original coloration during metamorphism.
3. Control of mineralization by lenses of originally grey pyritiferous rocks in a sequence of metamorphosed redbeds.
4. Zonal distribution of primary sulfide minerals around lenses of pyritiferous rock in the sequence: chalcopyrite-bornite-chalcocite, reflecting a drop in the concentration of the sulfide ion and an increase in copper content in the direction from originally grey rock to red rock.
5. A specific set of sulfides in primary ores (chalcopyrite-stoichiometric bornite-digenite-chalcocite-tetrahedral Cu_2S), reflecting relatively high-temperature conditions of formation (or transformation) of ore, corresponding to pT-conditions of the greenschist facies of metamorphism.
6. The presence of zones of silicification in the deposit controlling the region of occurrence of fine-grained rock (mudstones and siltstones) in the ore member and tentatively linked to processes of metamorphism.
7. Thorough reworking of the primary mineralization by supergene processes manifesting in secondary enrichment, oxidation, and leaching of sulfide ores.

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CARBONIFEROUS COAL-BEARING DEPOSITS, L'VOV-VOLYNA BASIN

V. F. Shul'ga

Formation analysis of Lower and Middle Carboniferous coal-bearing deposits of the L'vov – Volyna Basin involved discussed study methods, the composition, structure, coal content, paleogeographic conditions of accumulation of the formation and their subunits. Formation processes were studied and comparative analysis performed on Upper Paleozoic coal-bearing formations, located on shield slopes and in adjacent depressions. Prospective coal volumes were calculated. The efficiency of exploration and exploitation methods, as well as the coal mining techniques have improved. There is a need to study the single coal-bearing Group that occurs in the L'vov – Volyna and Lublin Basins.

The opening of the L'vov – Volyna coal basin for mining was a major accomplishment of domestic coal geologists in the post-war era. Characteristics of geological composition, coal content and quality, formation conditions of coal-bearing sediments in this basin have been discussed in numerous monographs [3-5, 8, 16, 17, 22, 23]. Complex facies analyses, performed recently on Lower and Middle Carboniferous coal-bearing deposits of the basin are discussed here.

Coal-bearing formations occur in the northwest corner of Ukraine in the L'vov Paleozoic basin and on the buried western slope of the Ukrainian Shield (USH) among the southwest margin of the East European Platform (EEP). The 5-50-km-wide distribution area of coal-bearing sediments extends for a distance of more than 120 km from the Polish border. From north, southward, the basin includes the Novovol'yn, Chervonograd, and the Southwest coal belts (Fig 1).

Carboniferous deposits overlie Middle and Upper Devonian terrigenous – carbonate formations across an erosional unconformity. The basal Carboniferous consists of maximum 120-m-thick Tournaisian – Middle Visian terrigenous – carbonate deposits (Khorevsk, Kulichkovsk, and Oleskovsk Formations). Paralic coal-bearing formations of the Lower Carboniferous Middle Visian and Serpukhovsk Stages, as well as the Middle Carboniferous Bashkirian Stage occur higher. Still higher in the section, Middle Jurassic terrigenous, variegated formations overlie a major unconformity. This, in turn, is overlain by Upper Jurassic evaporite – carbonate deposits.

Earlier the entire Carboniferous in the L'vov Paleozoic basin has been attributed to the coal-bearing formation [20, 21]. Based on the analysis of depositional condition of Carboniferous deposits, we now believe that only part of the section belongs to that unit. The higher, maximum 115-m-thick limestone and argillite beds (Oleskovsk Formation – "Uo") contain virtually no coal beds. This is significant genetically, as it establishes a transgressive – regressive trend in the formation of coal deposits in the L'vov – Volyna Basin. Two relatively thin members occur in the sequence, each characterized by specific composition, structure, coal content and formation conditions (Fig. 2). The lower member of low coal content, formed in regressive swampy – marine facies; the upper one of high coal content, formed in regressive – transgressive alluvial – lacustrine – swampy – lagoonal facies. Limestone bed N₃, or an included argillite bed with *Posidonia* and goniatite ammonoid fauna separates them.

The basin, in one direction extends to the buried slope of Ukrainian Shield, in the other, abuts on the Carpathian fold belt. Carboniferous rocks are inclined southwestward with a regional dip of 1-2 degrees. The linear fold pattern and widespread development of disjunctive faults were associated with tectonic movements of the crystalline basement (Fig 1).

RESEARCH METHODS

Experience in the Moscow and Donetsk Basins [9, 11, 24, 33] indicated that significantly detailed exploration in basins, involving composition, structure, coal content and depositional conditions of the formations, must be based on large-scale geologic mapping. A significant number of relatively evenly distributed data points is needed in the study of relatively thin section intervals of the same age.

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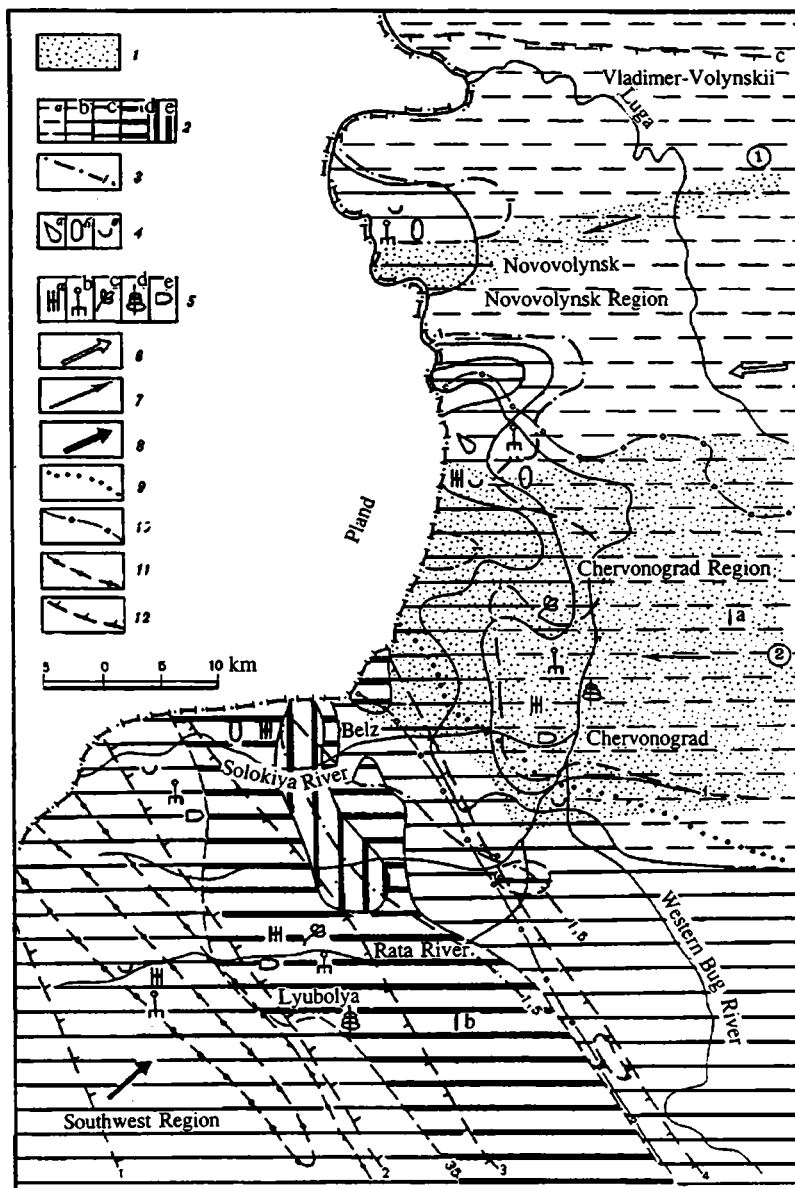
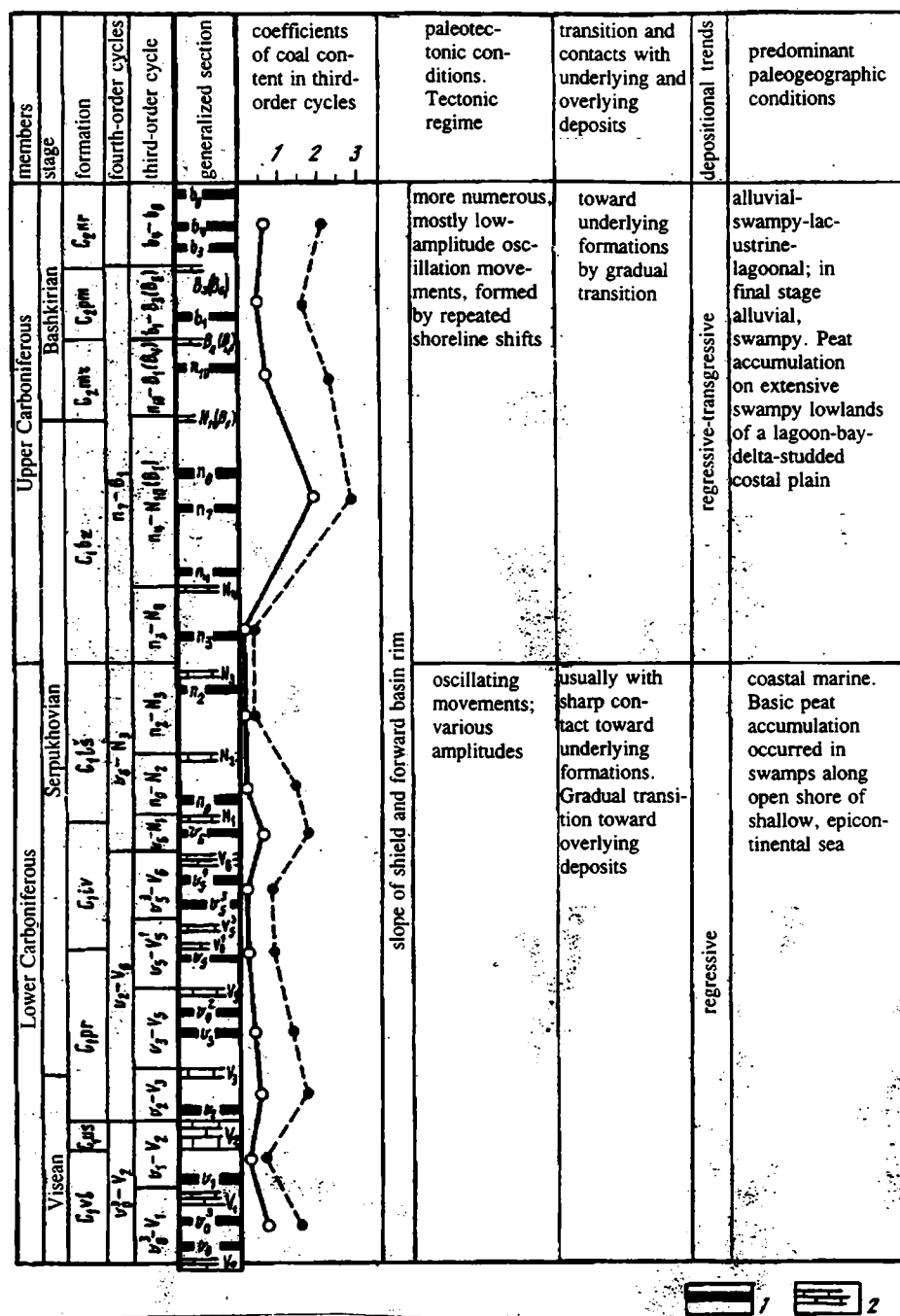


Fig. 1. Middle Visean-Late Serpukhovian lithological-paleogeographic map of the L'vov-Volyna Basin during deposition of lower coal-bearing member (lower Visean-to-late Serpukhovsk time). *Lithological-paleogeographic zones and subzones*: 1) shallow marine, lagoonal, coastal plain lakes and swampy zones: 1a) subzone of less active subsidence (5 m/km gradient), increased distribution of sandy-aleurolitic sediments in river mouth and floodplains (25-30%), lesser distribution of aleurolitic-clayey coal deposits in peat swamps (0.9%), coastal plain lakes and lagoons (20-25%); 1b) zone of more intensive subsidence, 6.55-7.5 m/km gradient. Increased clayey and coal deposits in peat swamps (1.2%), coastal lakes, lagoons (30-35%), lesser distribution of sandy-silty deposits in river mouths and floodplains (4-10%); 1) river mouths (sands and silts); 2) number of coal beds (a - maximum 20; b - 20-25; c - 25-30; d - 30-35, e - >35; 3) general coal index, %; 4) fauna (a - nonmarine bivalves; b - marine-lagoonal; c - marine); 5) flora (a - articulated *Calamites* stalks; b - club moss types; c - pteridosperms; d - ferns; e - *Cordaites*); 6) overall sediment transport direction; 7) stream flow direction; 8) direction of marine transgression; 9-11) boundaries (9 - lithological-paleogeographic subzone; 10) coal zone; 11) correlative member; 12) tectonic dislocation. Tectonic dislocations: 1) Rava-Russk fault; 2-4) thrust zones (2 - Nesterovsk; 3 - Butyn'-Khlevchansk; 4 - Belz-Milyatinsk); 5 - Northern (Vladimir-Volynsk) Fault. Paleostream systems (Arabic numbers in circles): 1) Luts'k; 2) Gorokhovsk-Rovens'k.



stream drainage network development	coal belt width	coal content	dominant facies conditions during coal accumulation	cyclicality	lithological composition	flora	fauna, "indications of life"	concretions
widespread	over 100 km	high total coal content index, 2%; commercial coal index: 0.9%. Maximum 40 discrete and relatively persistent coal beds of simple or somewhat complex composition. Eight beds mined	flooding and periodic flooding of wooded peat swamps along streams	well-expressed: fine and medium. In 1-20 order cycles alluvial-lacustrine-lagoonal, lacustrine-lagoon deposits, 8-20 m thick	sandstones, aleurolites, argillites, numerous coal beds and rare limestone beds	club moss types of articulated stems, pteridosperms	widespread nonmarine bivalves. Limited distribution of lagoonal, lagoonal-marine and marine faunas. Widespread distribution of tracks of mud-eating organisms and burrows of bottom-boring organisms. Widespread development of marine bivalves (<i>Posidona</i>) and gniatites at the bottom of the member	widespread distribution of nodules, layers. Increased FeCO_3 content
limited	over 200 km	relatively low coal content. Total coal-content index: 1%, commercial coal index: 0.1%. Maximum 45 non-persistent beds, mostly of simple composition. Four beds mined	flooded and intensively flooded peat swamps	more or less distinct. In 1-20 order cycles marine sediments pre-dominate. Thickness of cycles highly variable; range from 5-7 m to 25-30 m; 40-lithological composition	argillities, aleurolites, limestones, sandstones, numerous coal beds	articulated club moss types, pteridosperms	varied, mostly stenohaline fauna of the marine nearshore. Limited indications of biological activity by organisms	widespread concretion-bearing beds and lenses. Increased CaCO_3 and MgCO_3 content



TABLE 1. Facies and Facies Groups of Coal-bearing Formations

Facies (number of lithotypes)	Symbols
<i>Mostly Continental Group</i>	
Sandy and silty sediments; river mouths and alluvial stream floodplains (11)	Ur
Silty, sandy, and clayey river floodplain sediments (7)	Pr
Silty and sandy sediments (3)	Vz
Clayey and silty deposits of swampy nearshore marine coastal plains (4)	Nzp
Coal-clayey sediments of silted-up peat swamps and sapropel-rich lakes (1)	Btz
Coal deposits in peat swamps and sapropelitic lakes	Bt
<i>Transitional group</i>	
Clayey and silty deposits of marine coastal plain lakes highly brackish (decreased salinity) lagoons and bays (7)	Op
Clayey and silty lagoon deposits (3)	L
Silty and sandy deposits underwater bars (1)	Lp
<i>Marine group</i>	
Clayey and carbonate sediments, nearshore shallow marine areas (18)	Mpm
Clayey and carbonate deposits; offshore, relatively distant shallow offshore marine areas (2)	Mou
<i>Facies group of varying origins</i>	
Interlayered silty and clayey deposits of various basin facies types; coastal plain lakes, lagoons, bays, nearshore shallow water (3)	Vr

Our method consists of two basic parts. The first involves 1:25,000-scale mapping of coal beds, including lithostratigraphic formation profiles and large-scale map, depicting coal bed patterns. The second part includes 1:100,000-scale paleogeographic mapping. This is based on a complex of lithofacies, facies-cycle, facies-geochemistry, concretion, paleofacies, paleotectonic, and paleo-river network studies. Paleogeographic charts were compiled for three cycle stages that include coal beds (first-order depositional cycle): 1) end of regression, widespread swamp development in the region; 2) time of greatest marine regression and maximum development of peat swamps and sapropel-rich lakes; 3) end of maximum marine transgression before regression [9, 25, 33]. The charts indicate the paleogeographic regions, sediment transport directions, ancient stream flow directions, marine regressions—transgressions, distribution boundaries of cycles and coal beds, bed-wedgeout (zero thickness) contours, subdivision lines, contours of commercial coal bed thickness, syn- and post-peat bed displacement and erosion, rock structure features, fauna, flora, concretions, and other genetic characteristics.

The widespread distribution of ancient streams during formation of the coal deposits and their great impact on the formation of coal beds led to compilation of a 1:100,000-scale map that shows the distribution patterns of alluvial—deltaic deposits.

24 paleogeographic maps were compiled, each for a 10,000-12,000 year time-interval within the late Serpukhovsk-early Bashkir period. The maps were based on one hundred evenly distributed sections, representing a total of more than 35,000 linear meters. Twenty charts depicted sandy—silty deposit distribution patterns in river mouths, essential in reconstructing the ancient hydrographic systems.* Additionally, 1:100,000-scale maps were compiled to

*V. I. Garun, B. I. Lelik, S. S. Savenok, and E. A. Skovorodnikova compiled the sections.

TABLE 2. Coal-Bearing Carboniferous Formations

Members	Number of coal beds		Thickness of coal beds, m		Coeff. of coal content, in %	
	total	exploited	total	exploited	total	commercial
Upper	23	7	6	5.6	2	0.9
Lower	27	1	6	0.7	1	0.1
Total for formation	50	8	12	6.3	1.4	0.5

illustrate coal-bearing members. These depicted characteristics of lithological-paleogeographic, facies-geochemical, paleotectonic, coal bed-quality, coal content coefficients, and coal bed thickness.

FORMATION COMPOSITION AND STRUCTURE

Essentially terrigenous rocks; aleurolites (39%) and sandstones (19%) form the coal-bearing formations. Limestones represent 10%, coals, 1.4%. Limestones occur mostly in the lower member. Humic coals, represented by clarain-durain and durain-clarain lithotypes are the most widespread. Humic-sapropelic and sapropelic (boghead and cannel) coals are also common.

The formations have complex, polyfacies composition. Four facies groups occur that, in addition to coal, include 11 facies and 60 lithogenetic types (Table 1).

Lithogenetic types and facies display well defined patterns of changes in the section. Recurring paragenetic associations depositional cycles are recognizable. Unit cycles of the first order involve lithogenetic type and facies association that occupy areas of hundreds of square kilometers. They are of persistent thicknesses, at least 3-5 m thick. This indicates depositional conditions that alternated between deeper and shallower environments. Highly variable (5-30 m) marine cycles predominate in the lower coal-bearing member, while the upper member includes alluvial-lacustrine-lagoon, lacustrine-lagoon units, 8-20 m thick. Higher-order cycles consist of a combination of thinner cycles that alternate and display a regressive-to-transgressive, or reversed trend upward in the section. Sixth-order cycles in coal beds include two fifth-order cycles that correspond to coal-bearing members, five fourth-order cycles, fourteen third-order cycles, and over twenty second order cycles. More than one hundred first-order cycles are present (Figs. 2 and 3).

The formation thickens southwestward and exceeds 1300 m along the western basin boundary (Fig. 3). The following paleotectonic regions occur in this direction, based on changes in the gradient, and formation thickness: 1) the slope area of the Ukrainian Shield and the outer zone of the Paleozoic L'vov Basin (3-5 m/km gradient); 2) the interior basin area (maximum 6-7.5 m/km gradient). The boundary between these regions runs in a northwesterly direction, slightly north of the Belz-Milyatinsk thrust zone, southwest of Chervonograd (Figs. 1 and 4). The peripheral area of inner basin zone, between the Belz-Milyatinsk and Butyn'-Khlevchansk thrust zones, a 8-12-km-wide zone possesses lesser thickness gradients (2 m/km). Northwestward an isopach change occurs along an east-west trend in a maximum 20-km zone between Sokalem and Chervonograd.

The coal-bearing formation includes mostly continental (49%) and continental-to-marine transitional (32%), with lesser amounts of marine (21%) deposits.

The lower member is characterized by higher limestone (16%) and lesser sandstone (14%) content, wide distribution of marine facies (28%); the upper member by significant amounts of siltstones and sandstones (75%), minor amounts of limestones (0.2%), high coal content, widespread distribution of continental (57%) and transitional (33%) facies. The latter ones displayed zonal development. Two zones were established in the northwestern region. They trend north-east-southwest: Zone 1 - increased concentration of sandy-silty sediments in river floodplains of stream mouths; lesser amounts of coal deposits in peat swamps, carbonate-clay deposits of shallow seas (Slope of the Ukrainian Shield and periphery of the L'vov Basin); Zone 2 - increased amounts of coal-bearing deposits developed in peat swamps, shallow marine carbonate-clay deposits, and lesser volumes of sandy-silty beds in river mouths and floodplains (basin interior). A corresponding pattern-of facies distribution and thickness values, as well as spatial changes in these patterns emerged. This indicated the decisive role of tectonics, not only in coal thickness distribution but also in defining paleogeographic conditions. Our earlier identified paleotectonic area units are essentially structural facies zones.

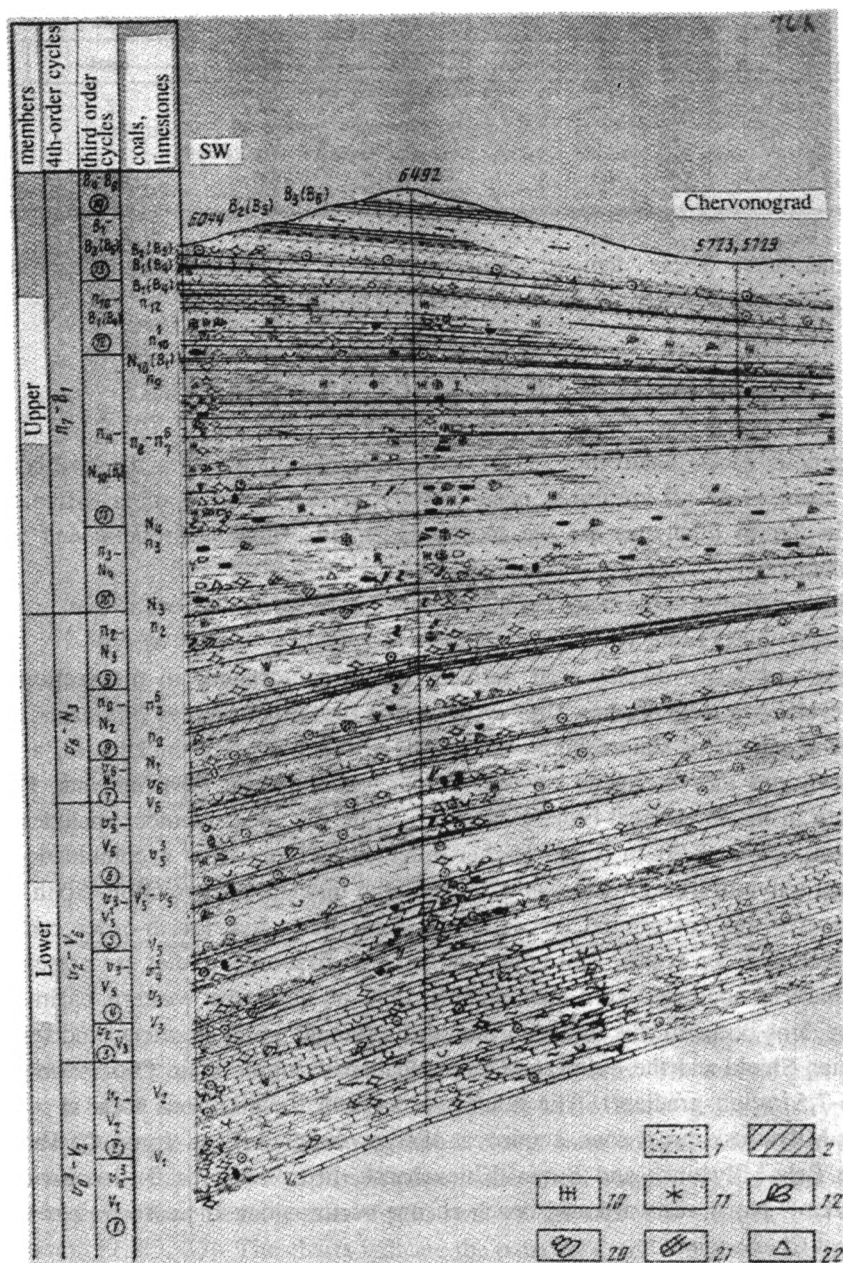
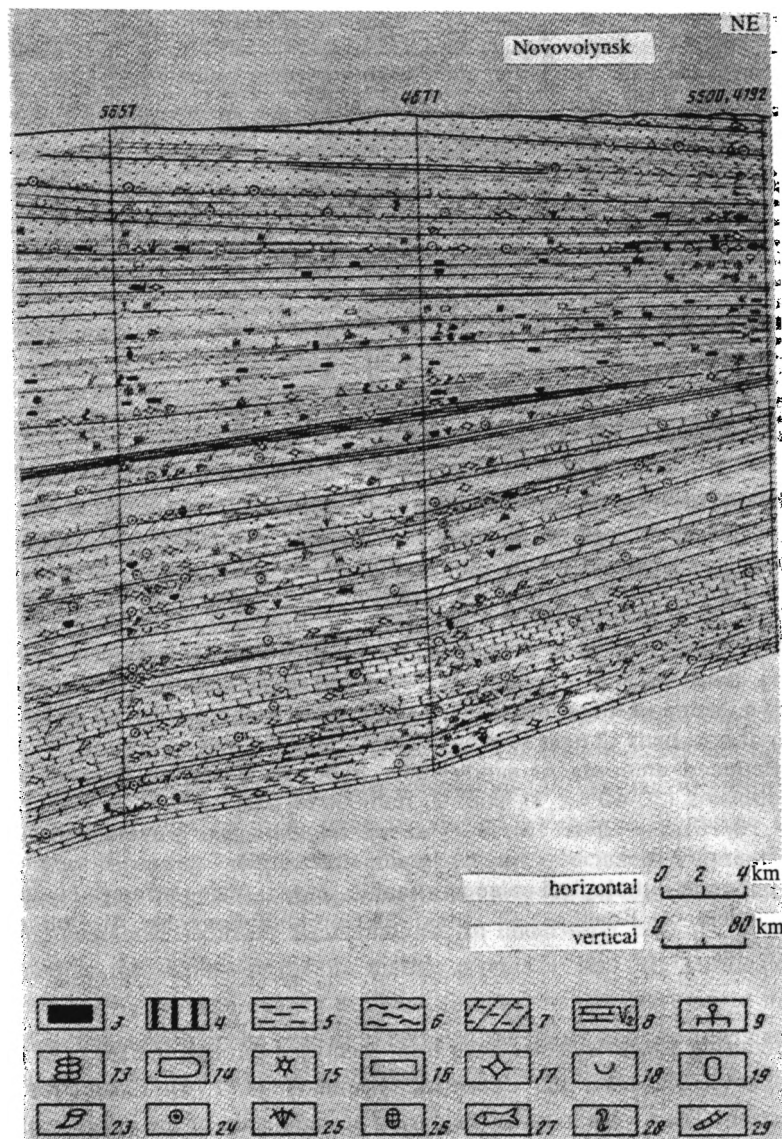


Fig. 3. Facies-paleoecological profile across coal-bearing formations. Comparison of the Novovolynsk-Chervonograd-Rava-Russkaya sections, 1-7) facies (1 – mouths, river floodplains; 2 – nonlapping basins; swampy coastal plains; 3 – peat swamps and sapropelitic lakes; 4 – silted-up peat swamps and sapropelitic lakes–coal-bearing argillites; 5 – coastal plain lake, lagoon, underwater; 6 – shallow sea; 7 – frequent facies alternation in filled-overgrown basin sequences; swampy coastal plain areas, nearshore lakes, lagoons); 8) limestone beds; 9-15) flora (9 – club moss-types; 10 – *Calamites*-types; 11 – *Sphenophyllia*; 12 – pteridosperms; 13 – ferns; 14 – *Cor-daites*; 15 – *Stigmara*); 16-29 – facies (16 – nonmarine bivalves, 17 – marine bi-valves; 18 – brachiopods; 19 – *Linqula*; 20 – gastropods; 21 – nautiloids; 22 – goniatites; 23 – corals; 24 – crinoids; 25 – bryozoans; 25 – trilobites; 27 – fishes; 28 – tracks of mud-eating organisms; 29 – burrows of boring organisms); Black: most euryhaline taxa.



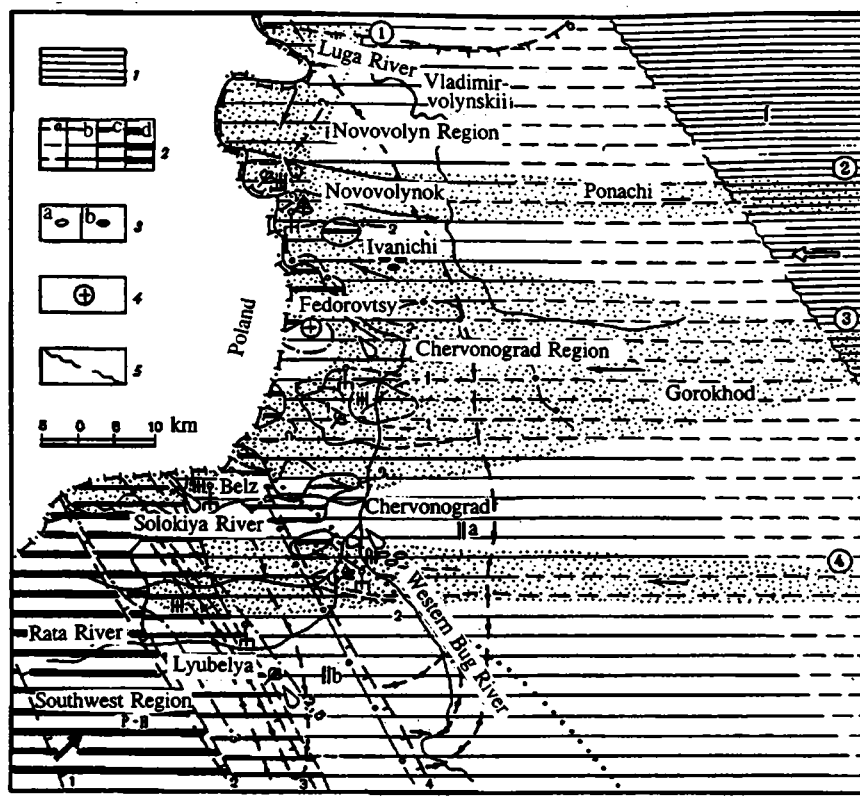


Fig. 4. Lithological-paleogeographic map of the L'vov-Volyna Basin at the time of the formation of the upper interval of member with high coal content of Late Serpukhovsk age. Lithological-paleogeographic zones and subzones: I) land (low, humid, eroded plain); II) swampy nearshore lowland, coastal plain lakes, lagoons, shallow sea: IIa) gently sloping subzone (2.5 m/km slope gradient). Increased distribution of sandy-silty deposits in mouth and low stream floodplains (40-50%); reduced distribution of coal-bearing peat swamp deposits (1.7%), carbonate-clayey shallow sea deposits (3-8%); IIb) subzone of more intensively raised areas (slope gradient 4 m/km), increased distribution of coal beds of peat swamps (2-3%); shallow marine carbonate-clay deposits (8-12%). Reduced amounts of sandy-silty deposits in stream mouths and lowland stream floodplains (30-40%); 1) land; 2) number of coal beds (a - maximum 5, b - 5-10; c - 10-15; d - over 15); 3 - exotic boulders and gravels in cal beds (a - mostly sedimentary rocks; b - mostly igneous rocks); 4 - areas of local uplifts; 5 - boundary between lithological-paleogeographic zones. Paleo-stream systems (Arabic figures in circles): 1) Kovel'sk; 2) Lutsk; 3) Gorokhov-Rovenski; 4) Dubnenski. Other symbols: as Fig. 1.

Four major, long-enduring stream systems have been identified through the study of the ancient stream network: the Kovel'sk, Lutsk, Gorokhovsk-Rovenski, and Dubnenski systems (Figs. 1, 4). Except for the first, all approximately parallel the east-west direction. They were most active during accumulation of the upper coal-bearing member.

The coal-bearing formation is characterized by widespread presence of club moss and articulated-stalked (mostly *Calamites*-type) plants; representing 46%, respectively 25% of the total macroflora. Lesser amounts of pteridosperms, ferns, and *Cordaite*-types also occurred. Hygrophytes were the most common; meso- and hydrophytes of lesser importance. The macroflora content initially increased upward in the section and achieved a maximum in the middle portion of the coal sequence, in the lower Lishnyanks Formation and middle and upper parts of the Buzhansk Formation. The concentration declines thereafter. The upper coal-bearing member differs from the lower one in the greater concentration of hydrophyte fossils (*Calamites*) and mesophytes, and in the smaller hygrophyte fossil content.

Bivalve molluscs are most common in these deposits (32%). Nonmarine forms (20%) and brachyopods (27%) predominated. Crinoids, corals, gastropods, nautiloids, ammonites, fishes, bryozoans and trilobites were less common. Foraminifers were widespread in the limestones. Attached benthic macrofaunal forms were very important (47%). Buried-infaunal (20%), immobile (20%), and mobile (4%) benthic forms played lesser roles. Tracks of mud-eating organism and animal burrows were widespread. Faunal abundance declines upward in the section, but the number of indications of biogenic activity increases. Nonmarine bivalves, fishes and ammonites represented an exception.

Four correlative bivalve molluscan biofacies have been identified within and in-between shallow and deep basin areas:

1) Numerous nonmarine sessile, and mobile, free-lying forms (*Najadites*, *Curvirmula*, *Anthraconaja*, *Carbonicola*) occur in river mouth, valley floodplain, coastal plain lakes, highly brackish lagoonal and bay facies. The assemblages include several-to-numerous taxa.

2) Relatively numerous attached sessile, free-lying, and infaunal benthic lagoonal forms (*Nuculavus*, *Palaeoneilo*, *Phestia*, pectinides, *Posidonia*). The biocoenoses contained mostly several-to-numerous taxa.

3) Numerous shallow marine, stenohaline, pseudoplanktonic infaunal forms (*Posidonia*, *Palaeoneilo*, *Nuculopsis*, *Phestia*, pectinides, *Edmondia*, *Selenomyalina*), represented by few-to-several-to-numerous taxa.

4) Few stenohaline, mostly infaunal forms (*Palaeoneilo*, *Nuculopsis*, pectinides) — in shallow marine zones at greater distances offshore. Biocoenoses contain few-to-numerous taxa. Similar facies associations were also established with regard to brachyopods.

With increasing offshore distances, fossil associations displayed the same trend in change. In continental, transitional and marine facies the respective concentrations were 68%, 26%, respectively 6%. Changes in the percentage values between facies groups were uneven. The first phytocenosis maximum (60%) corresponds to sandy-silty, and clayey river floodplain deposits, to lesser extent to swampy coastal plain facies. Biocoenoses of many taxa occur, with predominance of well preserved *Stigmaria*. Plant roots and appendages infiltrated rocks. *Calamites* stems, leaves, stalks, and stems of club moss types, pteridosperms, and ferns are also common. These were autochthonous and hypoallochthonous plant biocoenoses. A second plant biocoenosis maximum (25%) involves silty-clayey deposits of coastal plain lakes, greatly reduced salinity lagoons and bays. Allochthonous plant biocoenoses predominated, with abundant fossil types of both groups. Very rare and well preserved *Stigmarias* and their appendages occurred. Fossil plant appendages were common in the rocks. The number of phytocenoses declined with increasing distance offshore. Having been transported by wind, *Cordaitea* leaves were widespread in shallow nearshore marine argillites.

Large number of concretions occurred in the coal-bearing deposits. The average concretion content was 0.6%; the concentration content in third-order cycles ranged between 0.1-4.6%. Concretion-lenses and layers were most widespread (54%). Irregular concretion shapes were less common. Most concretions (68%) occur in argillites of the top of coal beds, less frequently in argillites and aleurolites of soils. Irregular shaped concretions dominate in the soils. V. P. Stasiv concluded that concretions in coal-bearing beds include numerous components and several mineral phases (siderite, sideroplesite, ankerite, and pistomesite). Concretion morphology and composition characteristics indicate a steady seaward trend. In swamp, lagoon, and nearshore sediment accumulations, in underwater bars the concretions generally were of irregular shapes. They formed in the shallow-nearshore marine zone. Ca- and Mg-carbonate concentrations increased seaward in the concretions, while Fe-carbonate content declined. Similar changes were noted in the chemical composition of the concretions downward in the section at the transition between the upper and the lower members with more marine characteristics. A similar, lateral (east-west) trend also occurred in seaward direction from the source area.

COAL CONTENT

88 coal beds and layers occur in the formation; 50 beds on the average. 30 of these are of commercial value [8]. Coal beds belong to the thin and very thin categories. Generally, they are 0.2-0.4 m thick. The average thickness of commercial coal beds was 70-90 cm. Occasionally, these reach 2.25 m. The largest number of coal beds occurs in the lower coal-bearing member. In general, the beds persist laterally and are of variable thickness. Simple-to-complex composition characterizes them. They become more metamorphosed toward the southwest in which direction the formation's thickness also increases. Coal categories range from long-flame- and gas-varieties in the Novovolyns'ka region, to bituminous and coke coals in the Southwest Region.

Data in Table 2 confirm an already known trend that extends upward in the section; the increase in the total- and commercially exploitable coal content [17]. This change is uneven and clearly of cyclic character (Fig. 2). The lower and less frequently the middle (regressive) intervals that consist of fourth-order cycles have higher total and commercial coal content, than the upper intervals. The middle interval of the p_7-v_1 cycle, correlative with the upper half of the Buzhansk Formation, displays the greatest coal content. This represents the most important coal unit of the Basin. It occurs in the upper coal-bearing member. As the coal content ratios indicate, a cyclic trend is present upward in the section. The number of coal beds also changes within a 50-m interval of the section.

Significant lateral changes in coal content also occur within the basin. This is a directional, zonal trend. As Figs. 1 and 3 indicate, from northeast to southwest the number of coal beds in the lower member increases, along with a corresponding increase in the thickness of coal-bearing beds. Maximum values occur in the northwest of the Southwest Region. Farther southwestward, the number of beds declines in the member. In general, low, significant, respectively, maximum coal content occur in parallel zones, positioned northeast to the southwest. A maximum of 30 and more coal beds occur in these zones. The zone boundaries in general trend northwestward. Their convoluted outlines were caused by reduced coal content in zones where east-west trending ancient stream systems had been active. These coal content trends have been confirmed by similar trends: change in the total coal content-index and total coal thickness values.

As in the lower coal-bearing member, an increase in coal content has also been observed from the northeast, toward the southwest, in the upper coal-bearing member (Fig. 4). Comparison of Figs. 1 and 4 indicates that the maximum coal content of the upper coal-bearing member, in reference to the lower member, had shifted southward by 10-15 km. This holds also true for the commercial coal content. Maximum values of commercial coal content displayed a southwestward shift in the section in the transition interval between the lower and upper coal bearing members. Second, the number of coal beds, their total thickness, and the coal content index initially increases, then declines from the northeast toward the southwest in the lower member. This fluctuation does not take place in the upper member. A correlative coal content decline was not observed in the upper member in the far west. In comparison with the lower member and due to the more intensive activity of ancient streams, the zone of significant coal content is discontinuous here. Increase in the number of coal beds toward the southwest resulted from their splitting into thinner units on a regional basis. Tectonic activity was responsible for this. Areas of peat accumulation subsided unevenly. New coal beds have also formed (Fig. 3).

The studied characteristics of vertical and lateral composition changes, composition and coal content of the formation is explained by its formation conditions (Fig. 2).

PALEOGEOGRAPHIC CONDITIONS OF FORMATION DEVELOPMENT

As the result of the Breton Phase of the Hercynian orogeny in the Paleozoic L'vov Basin that started in the early Lower Devonian, deposition of Middle and Upper Devonian terrigenous-evaporite-carbonate formations came to an end. Continental conditions prevailed for an extended time during the Tournaisian and early Middle Visean interval. Devonian units were eroded and terrigenous deposits formed. Continental conditions alternated with marine ones and limestones were deposited. Tectonic activity was accompanied by volcanism. Basalt covered the Nedzhvitsa-Lublin and the Kouka-Vlodavy-Lukova areas in the Lublin Basin [10, 35]. Erosional processes peneplained basin areas and weathering horizons formed. Thin terrigenous-carbonate sediments (Korevsk and Kulinckhovsk Formations) formed. Widespread accumulation of shallow water, mostly carbonate deposits (Oleskovsk Fm) took place during the subsequent major transgression in the L'vov-Volynsk and Lublin Basins [14, 17, 35].

The second half of the Middle Visean marked a new stage in the development of the L'vov Paleozoic Basin. The tectonic regime significantly changed, as did paleogeographic conditions of sediment accumulation, sediment composition and the makeup of the section. Contrasts between tectonic movements that involved large regions, diminished. They became more frequent; numerous regressions and transgressions took place against the background of an overall regressive trend. Relatively quick shoreline shifts resulted in frequently alternating depositional environments. Conditions that led to coal accumulation were the most important. During a rather prolonged period from the middle Visean, at the longest until the Middle Carboniferous, relatively thick terrigenous cyclic deposits of several facies formed in the basin. They included coal and limestone beds. The tectonic regime, the paleogeographic conditions of sediment and peat accumulation have undergone certain alterations that defined the composition and coal content characteristics in the section and served as criteria in identifying coal formations of the two members. Widespread distribution of marine facies and brief marine

regressions characterized formation conditions of the lower member. Certain areas emerged as land; peat deposits and ancient stream were of limited in extent. Stream source areas, occasionally inundated by the sea, occurred east of the Ukrainian Shield. The study of various fauna group indicated that the marine basins during accumulation of the lower coal-bearing member were closely interconnected not only with west European basins but also with east European seas [6, 7, 31, 32, 34].

The area of coal formations exceeded 200 km.* Peat surface and humid tropical climate in swampy regions on the coastal plains of open shallow seas. Southwest of Chervonograd in the L'vov Basin interior the southern basin area was more favorable for peat development (Fig. 1; Subzone 1b). Due to the more intensive subsidence of the region and associated terrigenous sedimentation, compensated deposition occurred in this part of the basin. This led to widespread swamp and peat development. The most favorable conditions occurred in the Southwest Region, between the Belz-Milyatinsk and Butyn'-Khlebchansk zones characterized by a more stable tectonic regime. Toward the southwest, in proportion to increasingly marine and transitional conditions, facies became gradually less favorable for peat development. Commercial coal deposits are only local, associated with the more elevated eastern and central basin areas and also with activity of the stream network that created a compensating system of sediment deposition (Bed u6, south of the Chervonograd region).

The generally regressive trend, caused by uplift of the Ukrainian Shield led to the southwestward retreat of the sea. Widespread continental sedimentation resulted and reduction in marine deposition. Conditions, favoring peat development became more frequent and involved increasing areas as they spread southwestward. This was associated with the earlier noted shift in zones of coal, including commercial coal bed concentrations in the transition belt between older and younger deposits. The coal beds split; became subdivided into thinner units.

A major transgression in middle Late Serpukhovsk times ended the beginning stage in the formation of coal-bearing deposits. Limestone N3 and thick argillites formed, marking a paleogeographic change from the dominantly marine and transitional (lagoonal and bay facies) of the lower coal-bearing members to continental and transitional facies (coastal plain lakes, reduced salinity lagoons and bays) of the higher members.

Accumulation of the upper member occurred during a further general uplift of the region and southwestward retreat of the sea. This process was accompanied by width reduction of the coal-bearing zone to 100-150 km, widespread development of continental conditions, rebirth of four stream systems, and transport of large amounts of continental deposits from the eroded zone. These events occurred in Buzhensk times on the west slope of the Ukrainian Shield, 50-60 km east of the present boundary of member's distribution area. Study of the composition of detrital deposits of ancient streams and also of exotic inclusions in the coal beds reflect the heterogeneity of rock composition in the source area. Two terrigenous mineral provinces were identified: a garnet-apatite-zircon (Novovol'ynsk and Chervonograd regions) and an apatite-zircon-garnet association (Southwest Region) [30]. The source matter of the coal accumulated in widespread peat swamp areas and sapropelic lakes that occurred in swampy lowlands of the lagoonal-bay, delta zones of the coastal plain. As during accumulation of the lower member, the effects of various tectonic movements resulted in two structural facies zones and associated areas of coal formation. They occurred on the slopes of the Ukrainian Shield and in the marginal basin zone, an area of weak-to-significant coal development. Due to regression, conditions favorable for peat bog development existed in all major areas toward the southwest. This explains the displaced position of the zone of maximum coal formation in the upper member in relation to that of the lower member.

The regressive sedimentation trend was interrupted by a series of significant marine transgressions in early Bashkir time. These cycles imparted a regressive-transgressive character to the upper coal-bearing member. As the upper part of the member has been totally eroded from the L'vov-Volyna Basin, it is apparent that deposition of alluvial-lacustrine-swamp-lagoonal facies in the upper coal-bearing member in this region and adjacent areas of the Lublin coal basin, was immediately followed by a marine transgression and the deposition of limey argillites and limestones of Unit S₁ [35]. Paralic deposits in the basin alternate with Westphalian lacustrine and lacustrine-fluvial facies upward in the section. In general, the lower paralic Lower and Middle Carboniferous coal-bearing formations of the L'vov-Volyna Basin and the upper, limnic Middle Carboniferous formations in the Lublin Basin are part of a single Carboniferous coal-bearing Group that until the post-Carboniferous erosional episodes covered a large region. The lower, low-coal member of the L'vov-Volyna Basin formed during the initial stage of L'vov Basin's oscillating subsidence. Its formation was related to

*In the Ukrainian part of the basin.

unfavorable conditions for coal deposition, typical of the initial coal accumulation interval. In this sense the member resembles the first member of alluvial-lacustrine-swamp facies and low coal content in the Jurassic coal-bearing formation of southern Siberia [19], the Ayachyaginsk Member in the Pechora Basin [12], the Ashlyariksk Formation of the Karaganda Basin [13], and the lower member of low coal content and swampy-marine facies in the Lower Carboniferous coal-bearing formation, Donets Basin [25]. More favorable conditions for commercial coal development existed in the L'vov-Volyna Basin during formation of the upper member of high coal content, and in the Lublin Basin during deposition of limnic coal-bearing sediments.

FORMATION ANALYSIS OF COAL DEPOSITS: SOLVING THEORETICAL AND PRACTICAL PROBLEMS

The above results of formation analysis of coal-bearing deposits in the basin broaden and more exactly define our assumptions on characteristics of the composition, structure, coal content, conditions of sediment and coal development in the initial stages of the formation of Upper Paleozoic regressive, regressive-transgressive paralic coal-bearing formation on the shield slopes and in adjacent basins. The extensive early and mid-Carboniferous coal-forming stages thus may be characterized in detail in the East European Platform area and the coal-bearing formations compared with those of other basins. Thus, comparison of Lower Carboniferous formations of the L'vov-Volyna and Donets Basins indicated significant similarities in lithofacies, cyclicity, coal content, and lateral and vertical changes in coal content, thickness, composition and structure of coal beds, as well as in the formation conditions of the coal-bearing members. Origins of general geological phenomena, associated with the sections may also be explained, along with their cyclic character of development [27]. This helps resolving specific coal exploration, development and exploitation problems. In particular, the southwestward shift of coal content maxima upward in the L'vov-Volyna Basin section indicates the potential of the outer L'vov Basin zone in the exploration for sufficiently thick commercial coal beds in the lower member. For the upper member the Basin's interior offers similar prospects.* Reconstruction of ancient stream networks did not only indicate the heterogeneous composition and distribution of the source area in the Ukrainian Shield but also serves as basis for predicting potential coal mining hazards. Such issues include the role of faults, roofs of mine workings in danger of collapse, ground water and natural gas that threaten mine workings. Measures may thus be taken in identifying and averting hazards in coal mines [28]. Formation analysis led to compilation of a guide and an atlas [26, 33]. This effort follows research, started by Botvinkina, Zhemchuzhnikov, Timofeev, Feofilova, and Yablokov [2]. In addition, these studies are essential in compiling atlases for the major coal basins [1, 18, 29]. Such visual documentation allows comparative analysis of the lithologic, lithogenetic types and facies of coal-bearing deposits in the L'vov-Volyna Basin with comparable units in other basins. Collection of lithologic, lithogenic, fauna, flora, and mineral composition data is needed to introduce the formation analysis method to exploration and mining operations for use by geological surveys and coal mining companies in the basin. This improves the performance of exploration and mine geologists. It improves drillcore and mine working documentation. In the final analysis, the method enhances the quality and efficiency of coal exploration and mining operations.

The study shows that formation analysis is most important in the study of coal deposits. The next task is a comparative coal bed analysis in the L'vov-Volyna and Lublin Basins. In combination with Polish geologists, a unified investigation would lead to clarification of the geological development of a single overall coal-bearing Group. This would help resolving a broad range of scientific, methodological and practical questions.

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FORMATIONS AND SEDIMENTARY ENVIRONMENTS WITHIN THE EARLY PALEOZOIC ACTIVE MARGIN (CHINGIZ RANGE, CENTRAL KAZAKHSTAN)

S. G. Samygin and T. N. Kheraskova

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Three different sedimentary environments are characterized on the basis of reconstructing the original relative positions of individual formations and formation series across the trend of Caledonian structures of the Chingiz Range and studying the composition and textural and facies characteristics of various Middle Cambrian – Middle Ordovician sedimentary and volcanogenic – sedimentary complexes: 1) basins with an oceanic crust; 2) island-arc uplifts, their slopes and foots; 3) tectonic syntaxis related to general complication of the synsedimentary structures.

As is known, contemporary active margins are distinguished by the greatest complexity of tectonic processes and diversity of sedimentary environments [1, 3, 4] reflecting changes in the geodynamic conditions of formation of various strata in the continent – ocean transition zone both in space and in time. To establish the complete picture of the characteristics of sedimentation in such environments, which are not always accessible to direct observation, it is important to study ancient analogues. The Caledonides of the Chingiz Range in the eastern part of Central Kazakhstan, which represented in the Early Paleozoic the active margin of the Kazakhstan – Siberian, or Central Asian, paleocean [2] with an exceptionally wide range of rocks of different facies types, can serve as one of them. This region has a very complex imbricate structure, an interpretation of which [7, 8, 11, 12] made it possible to carry out palinspastic reconstructions, to line up lateral series of synsedimentary structures characterized by certain material complexes, and to establish their nature and general regularities of development [9, 12].

Figure 1 shows the reconstructed relative position of individual formations and formation series across the trend of Caledonian structures of the Chingiz Range. It is seen from Fig. 1 that among the formations widespread in the Chingiz Caledonides are many volcanogenic ones. On the basis of the characteristics of the composition, degree of differentiation of the volcanic products, and character of eruptions they are divided into two main groups (for example, columns 4, 10, 12, 14–16, 33 and 3, 27–29, 31, 32a, 34, etc.) and correspond to the two main classes of ancient structures of Kazakhstan [5] – basins with an oceanic crust that had a melanocratic base of a gabbro – ultrabasic rock complex and with volcanic uplifts similar to present-day island – arc uplifts. The sedimentary and volcanogenic – sedimentary formations are often subordinate to volcanogenic, but are more informative and enable judging the facies environments within various parts of the main paleostructures. We will dwell on a brief description of the most representative concrete examples, having limited ourselves to Middle Cambrian and Upper cambrian – Middle Ordovician formation series. These series are fairly complete, are constructed relatively simply, and correspond to two large and basically identical cycles in the tectonic development of the region.

The entire diversity of Chingiz Middle Cambrian – Middle Ordovician sedimentary and volcanogenic – sedimentary formations can be assigned to three types of sedimentation characterizing three different geodynamic environments: 1) basins with an oceanic and suboceanic crust of the marginal-sea type; 2) island-arc uplifts, their slopes and foots; 3) tectonic syntaxes related to a general complication of the main structures and which occurred mainly in the vicinity of island-arc uplifts.

BASINS WITH OCEANIC CRUST

Most characteristic for such paleostructures are siliceous formations, which are paragenetically closely related to basalts of the ophiolite association, replace them up the section, and usually compose the sedimentary cover on them.

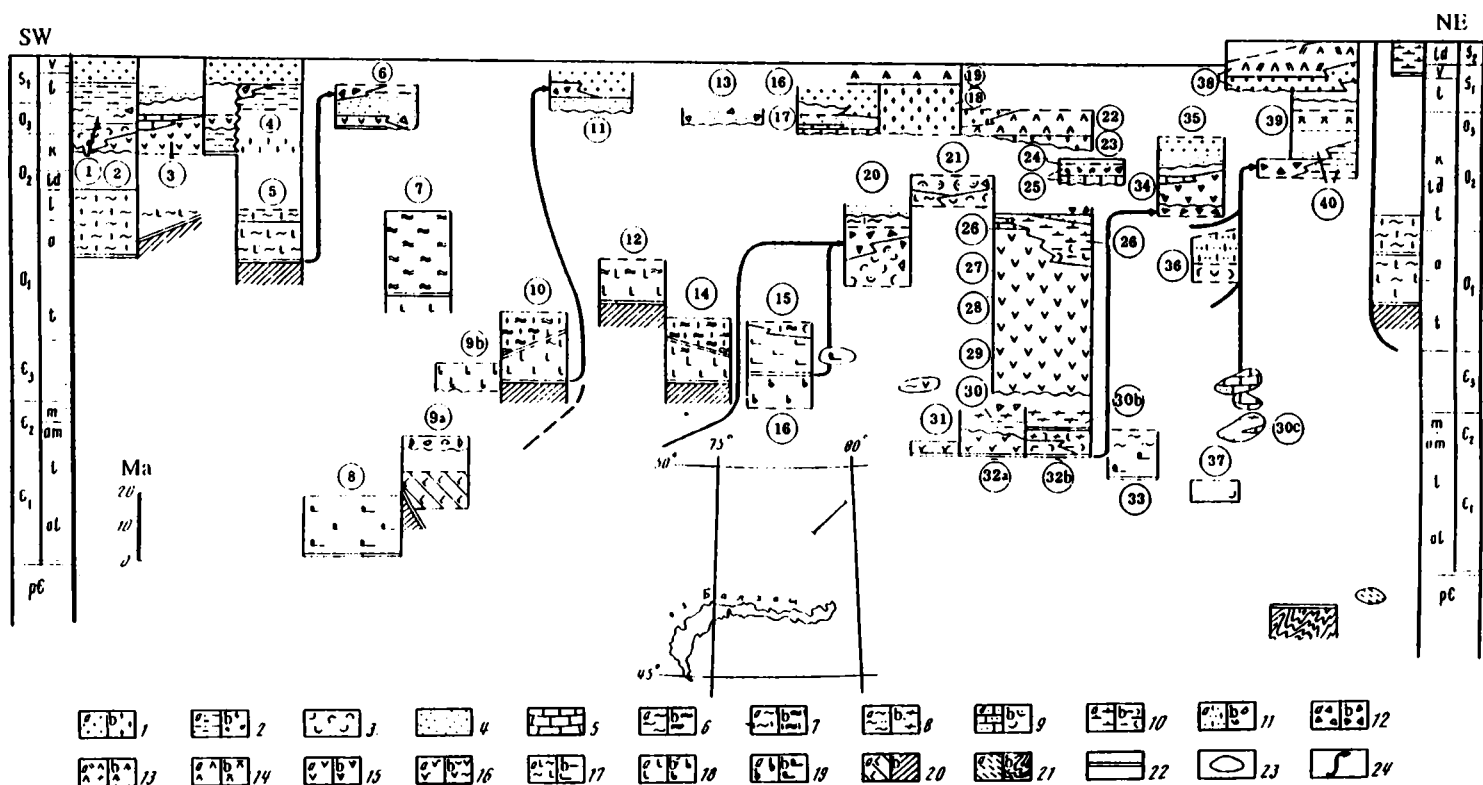


Fig. 1. Pre-Devonian formation series of the Chingiz Range (with elements of palinspastic reconstruction). 1-19) Formations: 1) lower molasse carbonate-terrigenous and terrigenous (a), carbonate-volcanogenic-terrigenous (b); 2) flyschoid terrigenous (a) and tephrogenic-terrigenous (b); 3) tephroturbidite; 4) terrigenous, carbonate-terrigenous nonrhythmic; 5) reef limestones; 6) jasper (a), phthanite-jasper (b); 7) tuffite-jasper (a), tuffite-siliceous (b); 8) terrigenous-tuffite-siliceous (a), tuffite-carbonate-siliceous (b); 9) terrigenous-carbonate (a), terrigenous-tuffite (b); 10) terrigenous-carbonate-siliceous (a), siliceous-tuffite (b); 11) siliceous-tuffite-terrigenous (a), carbonate-terrigenous-tephrogenic (b); 12) olistostrome with oval-block accumulations (a) and fragments of nappes (b); 13) andesite-basalt-dacite-rhyolite (a), trachyandesite-olivine-basalt (b); 14) trachybasalt-trachyandesite (a), trachyandesite (b); 15) basalt-andesite-dacite (a), andesite-dacite-rhyolite (b); 16) basalt-rhyolite (a), jasper-andesite-basalt (b); 17) jasper-basalt (a), carbonate-siliceous-basalt-andesite-basalt (b); 18) basalt (a), basalt-diabase (b); 19) alkali basalt (a), carbonate-basalt (b); 20) greenschists on basalts and tuffites (a), mixed gabbro-ultrabasic complex (b); 21) metapsammitic quartzites (a), greenschists, porphyroids, quartzites (b); 22) tectonic boundaries; 23) blocks in olistostromes and melange; 24) direction and time of overthrusting. Numerals in circles are suites and geographic ties of nameless strata: 1) Kuv; 2) Otyzbes Mts.; 3) Baban; 4) Balgashokin; 5) Kyzylkain Mts.; 6) upper reaches of the Samsa River; 7) Ushkzyzyl; 8) Balkybek; 9) region south of Kyzylzhal Ridge; 10) Tokai Mts.; 10-11) left bank region of Balkybek River west of Kyzylzhal Ridge; 12) Kyzylzhal Ridge; 13) south slope of Akchatau Range; 14) north of Tokai Mts. and Shapkaktas Mts.; 15) northeast of Koksengir Mts.; 15-16) Karaadyr Ridge south of Tokai Mts.; 17) Akdombak; 18) Al'peis; 19) Zhumak; 20) Shapkaktas Mts; 21) Shatkalan River; 22) Namas; 23) Taldyboi; 24) Sargaldak; 25) Bestamak; 26) Naiman; 27) Saryshokin; 28) Mamat; 29) Karakutui; 30 a, c) Chingiztau and Maidan; 30 b) Chingiztau; 31) Koksengir; 32 a) Zerbkyzyl; 32 b) Zerbkyzyl and Maidan; 33) Okpektin; 34) Abaev; 35) southeast of Zerbkyzyl Mts.; 36) right-bank region of Ayazuz River; 37) Edrei; 38) Donenzhal; 39) Zhartas; 40) Karagach. The location of the formation profile is shown in the inset.

Paragenesis of the basalt and siliceous formations is expressed especially distinctly in the Early and Middle Ordovician complexes. As a rule, they are strongly deformed tectonically and were preserved only fragmentarily as part of thrust sheets or individual erratic masses within the olistostrome strata. Nevertheless, the characteristics of the structure of the preserved siliceous sections and their comparison make it possible to rather definitely judge the conditions of sedimentation in various parts of the ancient basins.

The Arenigian–Llanvirnian phthanite–jasper formation (see Fig. 1, column 7) occurs on ferrobasalts. Its distinctive feature is the interbedding of dark-cherry jasper and phthanites. The following section (from the bottom up) is observed in the Ushkyzyl Mountains:

1. Thin-bedded dark-cherry jaspers with thin phthanite layers 5-7 m
2. Thin-bedded gray and black phthanites. Bedding is due to a different degree of enrichment of the thin beds in carbonaceous material. Phthanites contain few or no remains of radiolarians but are rich in conodonts. Among them are individual interlayers of light porcelainlike varieties distinguished by the presence of an admixture of fine ash material of an acid composition 70 m
3. Clay–siliceous shales slightly carbonaceous, pyrite-bearing with a large content of radiolarians, distinguished by thin-bedded and microfoliated structures 20 m
4. Dark-cherry radiolarian jaspers, manganous in the lower part. Jaspers have massive textures . . . 50 m
5. Sealing-wax radiolarian jaspers forming layers 3-20 cm thick separated by gouge. On the bedding surfaces are observed large nodular hieroglyphs of the gel-drop type as well as worm casts 70 m

The total apparent thickness of the formation is about 200 m

The presence of the remains of planktonic fauna in the phthanite–jasper formation and the occurrence of worm casts at the top suggest a relatively deep-water sedimentary environment. The presence of only very thin beds enriched with clay and fine ash material indicates remoteness of the section both from continental masses and from island–arc uplifts. Volcanic activity, apparently, promoted the accumulation of siliceous rocks and the flourishing of radiolarian plankton. Replacement of phthanites by jaspers up the section reflects, evidently, gradual expansion of the basin and improvement of the circulation of water masses.

The jasper–basalt formation at the end of the Early Ordovician (see Fig. 1, column 4) is represented by greenstone-altered basalts, most often without a distinct pillow structure. Thin interlayers and lenses of jaspers and well as of siliceous–ferruginous rocks are occasionally found among the volcanic rocks. The basalts are weakly differentiated, are closest in chemical composition to ferrobasalts of oceans and to low-magnesian basalts characteristic for certain marginal seas [1]. The siliceous–ferruginous formations found at the top of the basalt flows at the boundary with jaspers are unique rocks. They compose short (3-10 m) horizons and lenses with a thickness of 0.5-1 m and have a flow-banded structure with inclusions of basalt, jasper, and plagioclase fragments with a size up to 1-2 cm extended subparallel to the flow. Possibly [10], they are the product of underwater gas-hydrothermal activity related to basalt volcanism.

The jaspers form beds and lenses with a thickness of 0.3-3 m and have a cherry color, banded, lenticular-banded (due to diagenetic transformations), and sometimes breccia structure. In the latter case the jasper fragments and fragments of partially pulled-apart jasper layers are cemented by a siliceous–ferruginous gellike mass similar in composition to siliceous–ferruginous rocks which form intersecting veinlike bodies in the underlying basalts. Often noted in the jaspers is an admixture of fine ash material (hematitized glass, crystalloclasts) or layers 1-2 mm thick, rarely 1-2 cm, of graded-bedded or homogeneous fine-grained tephroids. The tephrogenic material, represented by quartz and albite, was introduced from a distant source by deep-water currents (microbedded varieties) or was the result of ash falls (layers with a homogeneous structure). The jaspers contain numerous radiolarian remains, occasionally sponges.

The interbedding of ferrobasalts and ore-bearing sediments permits speaking about the accumulation of the formation near an active tectonic zone (fault, spreading zone?) at a considerable distance from the occurrence of explosive volcanism (island arc?) that supplied to the basin fine-grained tephra of an acid composition. Possibly, the source of explosions with the course of time approached the region of basin sedimentation, as a result of which up the section began to appear horizons of fine-grained tephroids, and the jasper–basalt formation was replaced by a tuffite–jasper one.

The Upper Cambrian–Tremadocian tuffite–siliceous formation (see Fig. 1, columns 10 and 14) composes the cover of the essentially basalt formations. In its composition are developed dirty-green fine-grained siliceous tuffites and tuff pelites with acid pyroclastics, among which are noted thin horizons of green–red banded jaspers with infrequent radiolarians as well as of phthanites and carbonaceous-siliceous shales. In the Tokai Mts. the siliceous rocks contain less pyroclastics and include a richer complex of organic remains – radiolarians, sponge spicules, Inarticulata, and conodonts.

The absence of volcanomictic material, fine grain of the tephroids and tuffites, representing the remains of ash falls, noncalcareous character of the section, and occurrence of jaspers compel us to assume the development of the formation in relatively deep-water parts of the basin not so distant from the volcanic uplift as regions of the accumulation of jasper—basalt and, especially, phthanite—jasper formations. Probably, the tuffite—siliceous formation is an intermediate link between sections of central parts of the basin and formations of the foots and slopes of the island-arc uplifts examined below (Fig. 2).

ISLAND-ARC UPLIFTS

Sections characterizing this type of paleostructures have in Chingiz a far better preservation than the basin complexes and are distinguished by a greater diversity. In periods of activation of island-arc volcanism primarily pyroclastic—sedimentary formations were widespread, which laterally and vertically were closely related to differentiated volcanic complexes and formed mainly as a result of their erosion and due to pyroclastic material. Such formations usually compose continuous lateral series reflecting the degree of reworking and distance of transport of clastic rocks from centers of eruptions as well as the degree of their dilution by carbonate or siliceous material. Another group of formations accumulated under conditions of dying of volcanic arcs characteristic, for instance, for the Kyushu—Palau Ridge in the Philippines Sea [13], which lost its magmatic activity and experienced gradual sinking. Let us examine both situations mainly in the framework of two volcanic epochs.

The carbonate—terrigenous—tephrogenic formation of the Amga Stage of the Middle Cambrian (see Fig. 1, column 9a) occurs unconformably with erosion on Lower Cambrian mylonitized basalts that composed the marginal part of the Middle Cambrian island arc. The lower 400-500 m of the section of the formation is composed of green, tobacco-green polymictic sandstones, which are gradually replaced higher by a member about 500 m thick of volcanogenic—sedimentary rocks — tephroids, tuffstones, and tuffites with pyroclastics of a basic and acid composition. Short lenses and slide blocks of clastic and organogenic-clastic limestones are found sporadically over the entire section.

At the base of the formation is a horizon of breccia-conglomerate and coarse- and inequigranular sandstones consisting of fragments of plagioclase—actinolite—epidote shales and basalts of the underlying complex. Furthermore, volcanic rocks of an andesite and dacite composition, granitoids, granophyres, limestones as well as grains of quartz, plagioclase, and potassium feldspar are present in the sand fraction of breccia. Such a diverse composition of the clastic fraction of breccia-conglomerate indicates their basal character and formation due to the basement of the arc, partially metamorphosed, penetrated by granites and represented by a volcanic complex of primarily a basaltic composition with a small participation of intermediate and acid extrusive rocks. The poor roundness of the clastic material indicates its rapid burial.

The overlying sandstones form layers 3-10 cm thick distinguished by inversion and pendulum types of graded bedding with a change in granulometry from coarse-sand with an admixture of gravel to fine-sand. The characteristics of the textures permit relating the sandstones to sediments of immature turbidity flows that accumulated in the upper part of the paleoslope. At the top of the member is a horizon of calcarenites containing up to 30-50% grains of plagioclase, quartz, partially carbonatized volcanic rocks of an acid, intermediate, and basic composition, as well as of tuff pelites. Single grains of serpentinites, epidote, apatite, hornblende, and altered biotite are occasionally found. The presence in the calcarenites of rather large shell fragments, the absence of sorting of the clastic material, its good roundness, and small content of cement suggest shallow-water sedimentation conditions. In the upper part of this member is found an exotic block of limestones with Middle Cambrian trilobites. A train of limestone rubble is observed in the enclosing sandstones. Some limestone varieties have abundant carbonate cement containing inclusions of organogenic detritus. Evidently, the exotic block of limestones is close in age to the enclosing carbonate—terrigenous deposits and was only insignificantly moved downslope from a shallower zone of the island shelf. Erosion of these limestones in shallow water gave rise to the calcarenites, and collapses and slides furnished material to the upper part of the slope in the form of blocks and erratic masses accompanied by trains of rubble material.

In the middle part of the section of the formation pyroclastic material of a contrasting composition occurs in the sandstones, the quantity of which gradually increases upward. These are crystalloclasts of quartz and albite, epidote, titanomagnetite as well as lithic fragments represented by basalts and andesite—basalts and altered volcanic glass. The

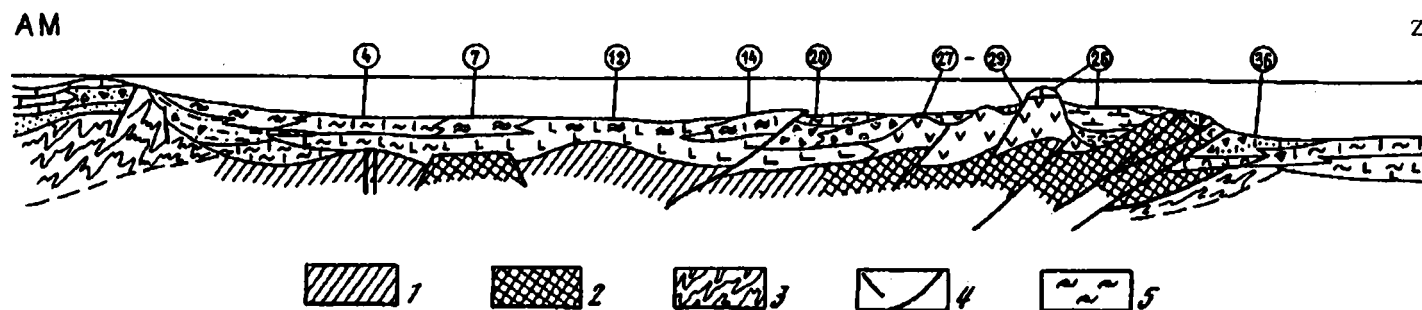


Fig. 2. Palinspastic profile for Early – start of Middle Ordovician. 1-3) Basement complexes: 1) melanocratic (oceanic crust); 2) syntaxis zones; 3) blocks with continental crust; 4) faults; 5) phthanite formation. See Fig. 1 for the other notations. AM) Aktau-Mointin massif; Z) Zaisan basin.

tephroids include an exotic block of pink limestones and calcareous gravelstones consisting of excellently rounded, in shallow water, grains of basalts, secondary quartzites from volcanic rocks of an acid composition, and tuffites. Distinct signs of a graded bedding characteristic of relatively mature tephroturbidites occurs in tephroids and tuffites in the upper part of the section, which indicates deepening of the sedimentary environment, possibly, gradual subsidence of part of the island-arc uplift. Infrequent radiolarians are found in the fine-grained tuffites. The rhythmicity of the tephroids is sporadically disturbed by the occurrence of tuff conglomerates consisting of clasts of rocks of an intermediate and acid composition, tuffites, sometimes quartzites. Organogenic limestones with clearly expressed sediment slide textures in an incompletely consolidated state complete the section of the formation. The limestones contain inclusions of scoria crusts, blocks of extrusive rocks of a basaltic composition, and interlayers enriched with polymictic clastic material.

Thus, the accumulation of the carbonate–terrigenous–tephrogenic formation occurred in the upper part of the slope of the island arc, probably on a terrace next to ledges composed of volcanic rocks of mainly a basaltic composition with a relatively small share of acid and intermediate rocks and also of ultrabasic rocks and granites. Here were carbonate shoals and banks, the clastic material from which as well as individual slide blocks periodically fell on the upper slope terrace with terrigenous sedimentation. With the course of time volcanic activity of a rhyolite–basalt composition with a high explosive index occurred on the island-arc uplift. The pyroclastic material was removed by autokinetic flows to the slope and deposited as tephroturbidites.

The tuffite–carbonate–siliceous formation of the *Maya Stage* (see Fig. 1, columns 30 a, b) replaces the volcanic rocks of the contrasting basalt–rhyolite formation participating in the structure of the Middle Cambrian island arc or with erosion covers the granites breaking through them. Dominant in the composition of the formation are green and variegated polymictic sandstones, tuffstones, tuff siltstones, and tuffites with acid pyroclastics gravitating toward its lower part, which are replaced higher and partially laterally by a thin ribbon or flyschlike alteration of argillaceous and siliceous tuffites, phthanitoids, and limestones. The sections are often disturbed by slide processes. The formation is distinguished by a distinct direction of facies changes vertically, reflecting gradual waning of volcanic activity and simultaneous deepening of the sedimentary environment. No less clearly displayed is lateral zonation, which is expressed in a change in the composition of the sedimentary clastics, depending on the composition of the underlying rocks, and in a change of the facies appearance of the sediments related to their deposition on various paleorelief features in the form of a series of terraces and horsts on the summit and slope of the sinking island arc.

At the base of more southern sections (Kol'denen River, see column 30 a) are developed large- and coarse-grained structureless sandstones consisting of clasts of rhyolites and quartz with a small admixture of clastic biotite and carbonate, which after 0.5-1 m are replaced by sandstones similar in composition but better sorted, having a gentle wavelike bedding of the coastal-marine type or coarse current bedding characteristic for alluvial–talus accumulations. Among the sandstones are found lenses of small-pebble conglomerates and gravelstones consists of well-rounded pebbles and gravels of granites, rhyolites, and siliceous tuffites. Layers with current bedding are enriched with rubble. Carbonate clastic material is also concentrated in individual lenses. The thickness of this member is about 10 m. Higher, the quantity of carbonate material gradually increases and the sandstones are replaced by a member of thin (1-2 cm) alternation of organogenic–detrital limestones with an admixture of crystalloclasts of quartz and calcareous tuffites with acid pyroclastics. Abundant remains of trilobites and gastropods are present in the limestones. The thickness of the member is 4-5 m. The more-upper parts of the section are reconstructed on the basis of individual olistoliths and olistoplakas occurring among clay siltstones not far from the place of original sedimentation (the thicknesses are only apparent):

- 1. Thin alternation of cherry tuff siltstones (layers 5-20 cm each), tuffstones, the red color of which is caused by a large amount of fragments of hematitized glass, and calcareous tuff siltstones with carbonate concretions (layers 2-3 cm each), often united into concretionary slabs 10-15 m
- 2. Pink calcarenites with an admixture of hematitized hyaloclastics distinguished by unpersistent bedding and numerous traces of roiling and rewashing of sediment 5-7 m
- 3. Gray micritic limestones 1-2 m
- 4. Thin rhythmic alternation of gray micritic limestones and microbedded phthanitoids with an admixture of fine acid hyaloclastics and remains of sponge spicules 3-5 m

Farther northeast along the Kol'denen River the section of the formation is distinguished by a coarse-fragmental composition, greater thickness, and absence of signs of carbonate content. Evidently, it accumulated lower along the slope of the paleoarc. At the base are small- and medium-pebble conglomerates of a dirty-green color with well-rounded pebbles of acid extrusive rocks, tuffites, and quartzites composing large wavy series (25-30 m), which are replaced higher by

sandstones with individual interlayers of puddingstones and gravelstones having a large wavy bedding of the coastal marine type (120-130 m). The upper part of the section here is also reconstructed from olistoplas. A member (0-15 m) of green tuffstones with interlayers and lenses of gravel of a volcanomictic composition and with fragments of slid cross-bedded coarse-grained calcarenites with an admixture of quartz crystalloclastics is distinguished. Then follow cherry microbedded tuffites with thin (2-3 cm) interlayers of siliceous tuffites with radiolarians, which after 2-3 m turn green and are replaced by a 15-20-m-thick member of olive-green, gray, and black thin-bedded phthanitoids and siliceous tuffs.

The northern sections of the tuffite-carbonate-siliceous formation (Zerbkyzyl Mts., column 30 b) are characterized by the deepest-water facies and more abundant admixture of pyroclastic material. Basal conglomerates and any signs of erosion are absent in the base here, the tephroturbidites of the lower-lying basalt-rhyolite formation (column 32 b) are conformably replaced by fine- and medium-grained acid tuffs and ashy calcarenites. Along the trend this horizon is replaced by landslide breccia consisting of a carbonate-tephrogenic matrix and clasts of limestones with trilobites. Higher again follow landslide breccia with a thickness of 10-15 m, but consisting of cherry and bluish-green siliceous tuffs. The upper part of the section is formed by thin-bedded fine-grained tuffites with signs of layerwise slumping and with inclusions of blocks of green-altered basalts from the basalt-rhyolite formation. The section is completed, as in the Kol'denen River region, with a member of green and gray siliceous tuff pelites, cherts, and phthanitoids. However, it differs by a substantially greater thickness (about 200 m) and contains at the bottom interlayers of turbidites, the coarse fraction in which is represented by clastic carbonate and the small and fine fraction by acid tephra.

It is necessary to emphasize once again the facies variability of the examined formation, the great effect on sedimentation of slide processes that occurred, apparently, during block subsidence of individual parts of the island-arc uplift. Another characteristic is the thin bedding of the deposits caused most probably not by the wide development of deposits of autokinetic flows but by the pulsating-discontinuous character of explosive and hydrothermal activity at the final stages of development of silicic volcanism, which was superposed on background terrigenous and carbonate sedimentation.

The new, most powerful and longest outbreak of island-arc volcanism is noted in Chingiz by a continuously differentiated *Upper Cambrian-Lower Ordovician basalt-andesite-dacite formation* (see Fig. 1, column 29-27). At the top it is represented by a complex, variable combination of lavas, agglomerate tuffs, tuff breccias with crystal tuffs, tephroids, and tuffites. Short lenses of organogenic-clastic limestones and calcarenites containing thin layers of carbonate-siliceous rocks with remains of sponge spicules are occasionally found.

Lavas and coarse-fragmental tuffs dominate in certain parts that represented earlier volcanic centers. Sometimes conducting channels are noted as a string of necks. Fragments of stratovolcano structures formed by the alternation of lithic tuffs with infrequent lava sheets are preserved more often. They are in complex facies relations with pyroclastic-sedimentary rocks of the tephroid and tuffite type, which are products of rewashing and redeposition of the pyroclastic material of central-type volcanoes under land and shallow-water as well as deeper-water conditions on the flanks of depressions that adjoined the volcanic structures, which rose as an island ridge. Thus, on the eastern limb of the Naiman syncline in the upper part of section 27 among tuffs are distinguished lithic, agglomerate, lithic-crystal, crystal-lithic, and ash varieties. They often contain inclusions of volcanic bombs with a size up to 1 m. The tuffs have an irregular texture, sometimes with signs of coarse stratification. Graded sorting of material is noted in varieties deposited through the water column during ash falls. More variegated colors, distinctly bedded structure, often with graded sorting of material are characteristic for the tephrogenic rocks; numerous traces of slides of unconsolidated sediments, intraformational erosion, and introduction and rolling of coarse material to fine are typical. The flow facies is noted by cross-bedded structures. The tephroids, representing products of the rewashing of tephra under shallow and nearshore conditions, are distinguished by a discontinuous, indistinct bedding, often by gradual changes of granularity, and by the presence of traces of eroded fine-grained material that did not form into layers. Sometimes among them are noted signs of coarse current bedding reflecting the activity of ephemeral streams.

The tephroids of deeper-water environments are sediments of immature turbidity flows characterized by a flysch-like rhythmic structure and graded sorting of material. The thickness of the rhythms is 0.7-2 m. At the base of each rhythm with a sharp contact and signs of erosion occurs coarse-grained tephroid, which is gradually replaced by medium-grained containing usually scraps of beds and "lumps" of tuff pelites. The rhythm is completed with fine-grained tephroid or tuff pelite. At the top the tuff pelites are often distinguished by a microhorizontal or current bedding. The microbedded varieties reflect, apparently, the activity of contour flows that redeposited the tephrogenic material, which arrived only in periods of explosions. The tephroids do not differ in composition from the tuffs and consist of fragments of the ground-mass of lavas of an intermediate, rarely acid composition, plagioclase crystalloclasts, sometimes with a subordinate amount

of pyroxene crystalloclasts, in silicic varieties — quartz, as well as small fragments of epidotized and chloritized volcanic glass, the maximum quantity of which is noted in fine-grained varieties. The ore mineral is often concentrated in the form of a heavy concentrate within individual small layers.

Two formation types are distinguished in the overlying rock complex (see Figs 1 and 2, column 26). The more western type (western limb of the Naiman syncline) is represented primarily by a terrigenous section of a *Late Arenigian terrigenous—carbonate—siliceous formation*. This is a 250-m stratum of greenish-gray sandstones, puddingstones, and gravelstones. At the bottom the sandstones are quartzose, higher arkosic. Horizontal- and wavy-bedded varieties alternate. The bedding is emphasized by enrichment of individual discontinuous layers in well-rounded gravel and pebbles or carbonaceous-clay material. Acid volcanic rocks as well as hornblende- and vein granites and unrounded fragments and blocks of organogenic limestones dominate in the pebbles. Along the trend a thin horizon of cross-bedded calcarenites is noted along with fragments of limestones. The coarseness of the clastic material decreases up the section, interlayers of siltstones appear, horizontal-bedded structures begin to dominate. The accumulation of this part of the formation occurred in a shallow-water environment of an island shelf after volcanic activity ceased in its limits, when processes of erosion of volcanic rocks and the granites breaking through them began to dominate.

Farther east (eastern limb of the Naiman syncline) the section is deeper-water, its thickness increases to 1000 m. Here the volcanogenic—sedimentary rocks of the basalt—andesite—dacite formation are gradually replaced by deposits of a *Middle Arenigian siliceous—tuffite formation* and younger *terrigenous—carbonate—siliceous formation* with a terrigenous component. At the bottom are observed light-gray thin-bedded tephroids, tuff pelites with acid pyroclastics, which are the sediments of currents of the contourites type, and medium-grained, primarily feldspathic and quartz—feldspathic sandstones with single granite grains, which were deposited most probably by grain flows. Then follow gray thin-bedded siliceous tuff pelites, calcareous tuffites with interlayers of phthanites containing thin-walled shells of brachiopods, graptolites, and conodonts. They are replaced by calcarenites with textures of turbidites and rare remains of redeposited trilobites. The thickness of the rhythms is 20-30 cm. The section ends with a horizon of radiolarian phthanites and graptolitic siliceous tuff pelites. Evidently, these rocks accumulated in a depression that occurred as a result of nonuniform block subsidence of the Late Cambrian—Early Ordovician island arc after cessation of intense volcanic activity and before the start of process of tectonic destruction with the formation of Llanvirnian olistostromes (see below).

The last example illustrates the characteristics of sedimentation in the island-arc basin that occurred in the final stage of Late-Cambrian—Middle Ordovician tectono-magmatic cycle on the back slope of the island arc composed of the andesite—dacite—rhyolite formation (see Fig. 1, column 34). This back-arc basin served as a trap for large volumes of clastic material. The Middle Ordovician formations — terrigenous—carbonate, tephrogenic—terrigenous, and terrigenous (see column 24-25) — filling it are characterized by considerable thickness, rhythmic flyschlike structure, abundance of volcanomictic material, and intense display of slide processes.

The terrigenous—carbonate formation begins with arkosic sandstones (40-50 m) deposited under nearshore, partially land conditions, which occur transgressively on Early Ordovician granodiorites or volcanic rocks of the Upper Cambrian—Lower Ordovician basalt—andesite—dacite formation. They are replaced up the section by a 180-m member of carbonate rocks composing the main part of the formation. The carbonate rocks form 0.5-0.7-m-thick layers differing in composition and textural characteristics. Interstratification of pelleted limestones, organogenic—clastic limestones with worm trails, micritic limestones, calcarenites, and carbonate breccia is observed. Organogenic—clastic limestones consist of fragments of shells of brachiopods and gastropods, columnals of crinoids, bryozoans. On the whole these are products of rewashing and redeposition in shallow water of sediments of bryozoan—brachiopod shoals of the inner part of the island shelf that topped the summits of nearby volcanic structures. Terrigenous clastics, presented by fragments of quartz and plagioclase crystals of silt size are present in various amounts in calcarenites. Intraclasts dominate among the carbonate clasts, pellets and fragments of oolites are present in a smaller quantity. Large wavy bedding of the coastal-marine type is developed in the calcarenites at the base of the beds, which higher is replaced by horizontal bedding. The breccias are the result of breaking the unconsolidated sediment in tectonically active periods. Apparently this same process caused the entry of the bulk of the intraclasts into the calcarenites. The topping member is transitional to the overlying flyschoid formation and is represented mainly by calcarenites of a lump structure, which occurred as a result of shaking up and slumping of the incompletely consolidated sediment. On the whole the accumulation of the terrigenous—carbonate formation occurred in the tectonically unstable environment of the island-arc shelf, mainly protected from the introduction of terrigenous material, during temporary cessation of volcanic activity. Probably, this was a shoal separated from the volcanic archipelago by a strait.

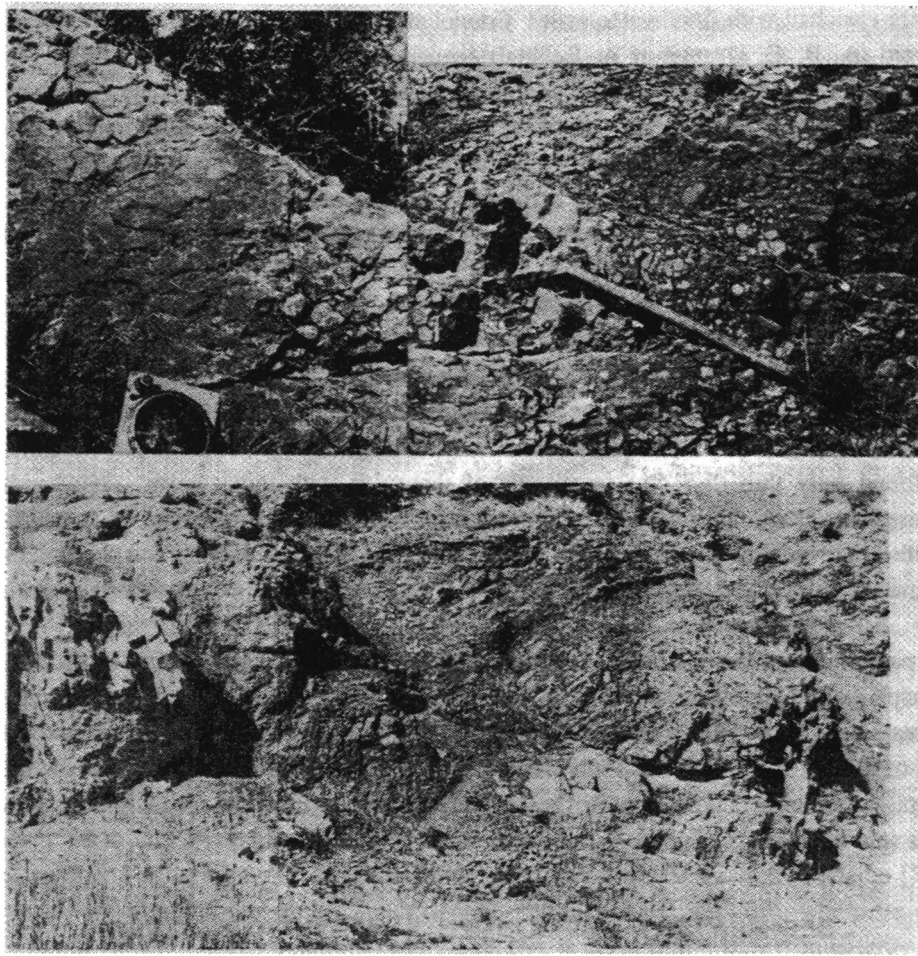


Fig. 3. Wildflysch at the base of the Middle Ordovician tephrogenic – terrigenous formation (right bank of the Shagan River). Various degrees of disintegration of carbonate rocks (a) and large limestone blocks (b) are seen.

The flyschoid tephrogenic – terrigenous formation (1300-2000 m) consists for dirty-green tuffstones, tuff gravelstones, tuffites, tephroids characterized by a distinct graded – rhythmic bedded and by the presence of thin horizons of coarse-fragmental mixtites. Both tephrogenic and volcanomictic material as well as carbonate is noted in the clasts. Fine-grained rocks contain abundant remains of graptolites.

Mixtites, or wildflysch [6], occur at the base of the formation directly above the calcarenites. At first gouge occurs between the fragments of the underlying calcarenites, higher the carbonate rock fragments begin to "float" in the fine-grained mixtite matrix (Fig. 3a) representing calcareous siltstone, which consists of small fragments of quartz, plagioclase and carbonate. Farther up the section the mixtite matrix becomes coarser, an admixture of pyroclastic material appears in it, the inclusions become larger and more diverse in composition (see Fig. 3b). Olistoplakas and olistoliths of organogenic – clastic limestones, calcarenites, arenaceous calcarenites from the underlying terrigenous – carbonate formation, and single olistoliths of hyaloclastites are present. Olistoplakas of carbonate rocks usually have a lumpy, conglomeratelike structure that occurred due to slumping of incompletely consolidated sediment. Apparently, the mixtites accumulated under conditions of an intensely subsided slope that formed at the site of the island-arc shelf. Slide processes had the main sediment-forming action. According to our data, nonuniform subsidence was related to waning of volcanic activity in the adjacent part of the Middle Ordovician island-arc uplift and to migration of volcanism northeastward.

The wildflysch is replaced higher by tephrogenic flysch. The thickness of the rhythms is 0.15-0.5 m. A sand element dominates. At the base of each rhythm with an erosion contact occurs coarse-grained sandstone, more rarely tuff gravelstone with pyroclastics of an intermediate composition and diverse volcanomictic clastics (basalts, andesites, rhyolites, plagioclase, quartz, pyroxene, limestones, tuff pelites with radiolarians). The coarseness of the material gradually

decreases, the rocks are enriched with fine hyaloclastics. Convolute bedding occurs in the tuffstones. The rhythm is topped by fine-grained tephroid (A, B, C, D units of A. Bouma's sequence). Some rhythms are separated by interlayers of tuff siltstones distinguished by thin horizontal or cross-bedding (sediments of contour currents). Scraps of beds of these thin-bedded tuff siltstones are usually present at the base of the overlying rhythm. Such a sequence is sporadically disrupted by individual horizons of mixtites having an underwater slide origin or representing sediments of pasty flows. The mixtites consist of fragments of tephrogenic rocks similar in composition to the matrix, small basalt blocks are found only occasionally among them. Furthermore, a concave lens with coarse-grained cementless tephroid, which contains scraps of beds of tuff pelites roiled and eroded in an unconsolidated state, is found. Apparently, this is a facies of underwater channels. On the whole, the lower part of the formation can be interpreted as deposits of the upper (proximal) part of a submarine fan of clastic material located at the bottom of the slope and foot of the Middle Ordovician island arc.

In the middle part of the section the character of the graded—rhythmic bedding changes. The coarse- and medium-grained rhythms similar to those described above or with omission of the C and D units are layered by rather thick (7-8 m) members of homogeneous structureless sediments that occurred during settling of ash through the water column. Among the tuff pelites are found only thin (5-7 cm) interlayers of tephroids with thin current cross-bedding. Coarse-grained rhythms erode the fine-grained sediments, and sometimes are enclosed in them. Apparently, layers of fine-grained tuffs and tuff pelites interrupted by current sediments characterize background pelagic sedimentation, only sporadically interrupted by the activity of immature turbidity flows which carried coarse tephrogenic and volcanomictic material into the gradually deepened basin.

In the upper part of the section of the tephrogenic—terrigenous formation the number and thickness of the coarse rhythms again increase, individual olistoliths of tuffs and tephroturbidites of an intermediate composition appear. This indicates a new outbreak of volcanic and tectonic activity within the limits of the adjacent island arc, which led to intensification of the activity of autokinetic flows and slide processes as well as to the formation of a new lobe of the submarine fan of tephrogenic and volcanomictic material in the back-arc basin. In the final stage the basin was filled only with terrigenous flyschoids (up to 1000 m).

SEDIMENTARY ENVIRONMENTS IN SYNTAXIS ZONES

Preceding investigations [7, 8, 11] proved that the majority of the examined formations presently compose a series of tectonic sheets separated, as a rule, by thick olistostrome strata, the accumulation of which occurred synchronously with nappe formation (see Fig. 1). A study of the structure of olistostromes, their composition, and parageneses with other formations makes it possible to interpret many characteristics of this process that played an important role in the evolution of island-arc uplifts and their adjacent basins. In Chingiz the formation of olistostromes occurred repeatedly and in various environments, which is clearly seen by two examples.

The uppermost Lower Ordovician olistostrome formation (see Figs. 1 and 2, column 20) occurred, according to our notions, as a result of syntaxis on the margin of the Late Cambrian—Early Ordovician basin. In the composition of the olistostrome (apparent thickness not less than 1000 m) is found a relatively fine-fragmental matrix in which are included olistoliths, olistoplakas, and large tectonic erratic masses (200 × 1000 m) of various older rocks. In the lower part is observed an alternation of bluish-green siliceous tuffites, dirty-green tuffstones and argillaceous tuffs of an intermediate composition, as well as poorly sorted gravelstones consisting of fragments of siliceous tuffites and tuffs. Sporadically found are interlayers of breccia-conglomerates not persistent along the trend with fragments of the enclosing tuffstones, more rarely of quartzites and limestones. In all rocks are noted traces of slumping and breaking of the unconsolidated sediment and inclusions measuring 0.5-3 m of blocks of quartzites and limestones with Upper Cambrian fauna, as well as hyaloclastites. Trains of smaller subangular clasts of the same composition and more intense slide structures are observed in the matrix around the blocks. Apparently, slumping of the sediments provoked movement of large blocks.

Higher up the section fine-grained tuffites, tuff siltstones, and tuff argillites with inclusions of rubble and clasts (mixtites, or diamictites) dominate in the matrix of the olistostrome. Here are found the largest sheets and olistoplakas. Blocks and olistoplakas disappear in the upper part of the formation, and in the matrix appear more distinct signs of sorting of the material, more persistent undisturbed bedding. However, here are found individual horizons of clay mixtites with inclusions of fine rubble and siliceous tuffs. Crystalloclasts of albitized plagioclase, in a smaller amount potassium feldspar and quartz, dominate in the composition of the clastic material. Lithic fragments are present more rarely and are

represented by tuff pelites, siliceous tuffs, basalts, and volcanic rocks of an intermediate composition. Siliceous rocks, granites, and volcanic rocks of an acid composition are found in single grains. On the whole, volcanomictic material dominates, a large part of it was introduced from an island volcanic uplift located farther northeast (see Figs. 1 and 2, column 29-27). The fragments and scraps of the siliceous tuff and tuff pelite beds are the result of slumping of the background sediments and their "ripping" by the high-density gravitational flows.

The olistoliths, olistoplas, and sheets included in the matrix are represented by three rock associations.

1. Aphyric basalts, siliceous tuffs, jaspers, phthanites, and secondary quartzites from them from the Upper Cambrian—Tremadocian basalt, siliceous—basalt, and tuffite—siliceous formations (see column 14).
2. Hyalobasalts, basalts, rhyolites, crinoid clastic limestones with an admixture of basic and acid pyroclastics containing the remains of Upper Cambrian trilobites. This association is referred to the carbonate—siliceous—basalt—andesite—basalt and carbonate—basalt formations of an analogous age (see column 15).
3. Andesites with large phenocrysts of zonal plagioclase, pyroxene, and hornblende as well as their tuffs, tuff pelites, limestones with an admixture of ash, similar on the whole to rocks of the island-arc basalt—andesite—dacite formation (see column 29-27).

The composition of the matrix and inclusions foreign for the section shows that its formation occurred not far from the foot of the Late Cambrian—Early Ordovician island arc during waning of volcanic activity within its limits, which manifested itself only by periodic ejections of tephra (compare with section 26). The large-block island-arc material has a collapse—slide origin. The participation in the structure of the olistostrome of sheets and erratic masses composed by marginal-basin facies is undoubtedly related to the formation of nappes. Expansion of the sea floor of the interior parts of the marginal basin — back with respect to the island arc (see Fig. 2) — with corresponding rejuvenation of its crust (spreading) could serve as the most possible cause of nappe formation. This is indicated by the symmetric arrangement of fragments of older basalt formations relative to younger ones (see Fig. 1, columns 10, 12, and 14).

The olistostrome formation of the start of the Middle Ordovician (see Fig. 1, base of column 34) has a thickness greater than 1500 m and accumulated in front of the island arc as a result of overthrusting and destruction of the island-arc complexes spanning an age interval from the Middle Cambrian to the Early Ordovician [7, 12]. The matrix of the olistostrome is relatively fine-grained, as a rule, abundant, although there are parts when the olistoliths adjoin one another practically without cement. In the apparent base the matrix is represented by greenish-gray coarse-grained sandstones, gravelstones, conglomerates with scattered, well-round pebbles of tuffites, limestones, and hornblende granites. The rocks have cross-bedded structures of the coastal-marine type, which are replaced higher by horizontal-bedded structures corresponding to progressive deepening of the sedimentary environment. In the more-upper parts of the section the matrix consists of tobacco-green homogeneous siltstones containing lenses and unpersistent interlayers of fine-grained polymictic sandstones. Sandstones dominate in some parts, in addition graded sorting of material characteristic for sediments of autokinetic flows occurs in them. Near olistoliths the matrix is enriched with rubble of rocks composing these inclusions. Albite and quartz dominate in the clastic fraction of the sandstones and siltstones, intermediate and acid volcanic rocks, tuffites and siliceous tuffs are present in a smaller quantity, very rarely metamorphic rocks. The clastic fraction is formed mainly due to relatively mature volcanomictic material that occurred as a result of erosion and transport of the products of disintegration of the differentiated volcanic series. Edaphogenic tuffite material had smaller significance.

The olistoliths and olistoplas are also represented by fragments of differentiated formations of the island-arc type: Middle Cambrian basalt—rhyolite (see Fig. 1, columns 31, 32) and younger basalt—andesite—dacite (column 29-27). Furthermore, individual olistoliths of pyroclastic—sedimentary formations, which composed the cover of the volcanic uplifts that lost their activity, are found (columns 30 and 26). All aforementioned erratic masses occurred ahead of the overthrusting front, which with the course of time moved from the southwest to the northeast [7, 8]. Nappe formation began with thrusting of the island-arc uplift that gradually subsided and the formation as a result of this of a mixtite stratum that formed in a shallow-water environment (Fig. 4). Subsequently, the nappes advanced into a relatively deeper-water environment of the former slope of the island arc, as a result of which considerable complication of the structure occurred with general thickening of the crust, which served as the "foundation" for the new volcanogenic complex (see Fig. 1, column 34).

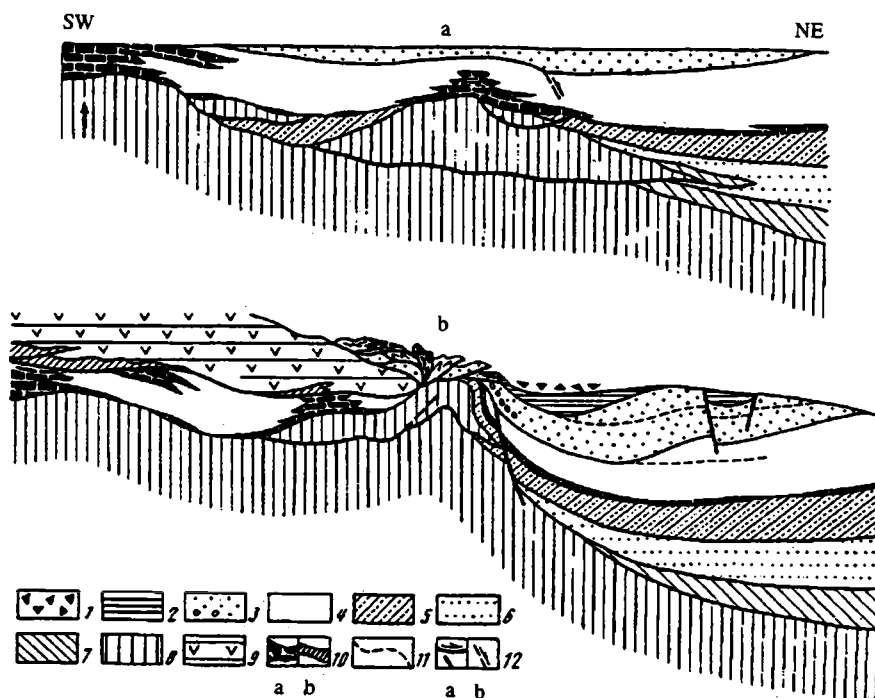


Fig. 4. Paleoprofiles illustrating subsidence of the slope of Late Cambrian – Early Ordovician island arc (a) and start of olistostrome formation (b) for the example of the Naiman syncline: 1) gravel – block mixtures, boulder – pebble conglomerates and breccia-conglomerates (O_21_{1-2}); 2 and 3) terrigenous formation (O_21_1): 2) polymictic gravelstones, breccia-conglomerates, inequigranular sandstones, 3) quartz – plagioclase sandstones, quartzose at the bottom, here and there with polymictic conglomerates and puddingstones; 4-6) terrigenous – carbonate – siliceous and siliceous – tuffite formation (O_1a_{2-2}): 4) dark thin-platy siliceous, sometime calcareous tuffites with interlayers of acid tephroids, 5) carbonaceous – argillaceous – siliceous shales and phthanites with tuffite horizons, 6) siliceous-chloritic silty tuffites with thin interlayers of arkose and siliceous clastic sandstones; 7 and 8) tephroid facies (O_1a_{1-2}) of basalt – andesite – dacite formation: 7) tephroids and tuffites with beds of conglomerates, tuff siltstones, tuffites, tuffs of intermediate and acid compositions, 8) heteroclastic tephroturbidites, tuffs, volcanic breccia of intermediate composition and their erosion products, siliceous tuffites; 9) andesitic tuffs, volcanic breccia, more rarely lavas, interlayers and lenses of volcanomictic sandstones and tephroids ϵ_3^{2-3} of the basalt – andesite – dacite formation; 10) limestones, calcarenites O_1a_3 (a) and ϵ_2^3 (b); 11) marker beds; 12) synsedimentation (a) and incipient (b) faults.

* * *

Summarizing the material presented, we must emphasize once again that in the Caledonides of the Chingiz Range three main types of sedimentation characterizing three main geodynamic environments dominated during the Middle Cambrian – Middle Ordovician. Pelagic, primarily siliceous condensed sedimentation virtually without the entry of any clastic allochthonous material dominated in the *first* type. Planktonic fauna and noncalcareous sediments are typical. Apparently, this type of sedimentation was characteristic for rather extensive and relatively deepwater marginal-sea basins. Facies persistence of the sediments and stable thicknesses of individual members and layers are characteristic.

The *second* type is completely opposite to the first: sedimentation occurred in an environment of intense, often avalanche entry of clastic material of a tephrogenic and volcanomictic origin. Sediments of various gravitational flows and slide accumulations of shallow-water limestones are typical. Facies replacements, contrasting changes in thicknesses, replacement of relatively shallow-water sediments with benthic fauna by deeper-water turbidites both laterally and vertically are noted ubiquitously. This type of sedimentation fixes the contrasting relief of the island-arc uplifts, their slopes and foots both during periods of outbreaks of volcanic activity and in relatively quiet epochs.

The *third* type of sedimentation is more complex. It was realized in an environment of convergence of the first to zones of sedimentation with a reduction of the space that separated them as a consequence of syntaxis. As a result olistostrome complexes became widespread. In some of them the matrix was sediments characteristic for the lower parts and foots of island-arc slopes (second type of sedimentation), and deposits of basin environments (first type) dominate among the olistoliths. Cases of opposite relations are also known. Other olistostromes occurred as a result of mixing of island-arc formations of different facies (and different ages) during migration of the island arcs themselves.

On the whole, virtually all sedimentary complexes characteristic for present-day active margins are developed in Chingiz. However, they are often intensely dislocated, and their original location is greatly veiled or changed repeatedly by superposed tectonic processes. As is seen from Fig. 1, their type of sedimentation related to one of such processes occurred in Chingiz only at the end of the Early Ordovician, and its role increased with the course of time. This reflects, evidently, stages of gradual complication of the structure of the Chingiz Caledonides and the occurrence of an extensive imbricate-nappe structure at the place of the former active margin of the Central Asian paleocean.

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INITIAL METAMORPHISM OF ROCKS OF THE VERKHoyANSK TERRIGENOUS COMPLEX (SOUTHERN VERKHoyANSK REGION)

I. M. Simanovich and N. G. Andriyanov

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A broad spectrum of postdepositional transformations of terrigenous strata of the Verkhoyansk complex – from deep-seated epigenesis to metamorphism of the epidote–amphibolite facies – was established within the South Verkhoyansk synclinorium. There exists distinct tectonic control of the distribution of postdepositional transformation in the area: the highest metamorphism is characteristic for zones of influence of concealed deep faults of a north–northeast strike. Zones of relatively high metamorphism are intersecting with respect to stratigraphic boundaries and fold structures.

The Verkhoyansk fold zone (together with the Vilyui syncline) is a classical object for studying postdepositional zonation in thick strata of terrigenous rocks. A. G. Kossovskaya considers that for studying postdepositional transformations "...the complex of terrigenous deposits of the southeastern part of the Vilyui depression and western Verkhoyansk region is a quite good object. A general facies–genetic series of formations is observed here both in the platform and geosynclinal zone, and the primary composition of the clastic material is identical" [3, p. 169]. This conclusion can be applied also to the Permo-Carboniferous Verkhoyansk complex of the northern and southern Verkhoyansk region, represented by thick homogeneous rhythmic terrigenous strata. In 1955 A. G. Kossovskaya and V. D. Shutov published the first scheme of postdepositional zonation of the Vilyui syncline and western Verkhoyansk region [4]. On a zonation map published later [5] it can be seen that the zone of initial epigenesis* encompasses Cretaceous and Jurassic strata of the Vilyui syncline, the zone of deep-seated epigenesis is characteristic for Triassic–Jurassic strata of the western part of the Verkhoyansk fold zone (VFZ), the zone of metagenesis is confined to Permian deposits of the axial part of the VFZ. A fourth zone (metagenesis – spiniform textures) is developed in Lower Permian rocks as spots and is also confined to the axial part of the VFZ (Table 1).

In 1973 N. G. Andriyanov [1] established that within the South Verkhoyansk synclinorium metamorphism of biotite and even staurolite metamorphic grades is developed as narrow zone on rocks of the terrigenous complex (see Table 1). This conclusion is confirmed in [6]. Later, for the example of the northern Verkhoyansk region (Kharaulakh), O. V. Yapaskurt [8] also established a similar postdepositional zonation (see Table 1). Thus, within the VFZ and Vilyui syncline lithogenesis of the terrigenous strata is fairly well studied and on the whole one can speak about a unique typical section with a vast range of intensity of postdepositional transformations of the terrigenous strata.

In this connection it is of interest and important to examine the character of transition from deep-seated epigenesis to metamorphism of terrigenous strata for the example of the southern Verkhoyansk region.

GEOLOGIC STRUCTURE

As shown in [8], the accumulation of enormous masses of polymictic terrigenous rocks, evidently, as a result of avalanche sedimentation in delta fans, occurred under tectonically mobile conditions of the Verkhoyansk miogeosyncline (passive continental margin), which predetermined the flyschoid appearance of the strata. It was already noted that on the whole the surprisingly uniform strata of the complex are represented by a rhythmic alternation of sandstones, siltstones, and argillites with the predominance of various varieties at different stratigraphic levels. The presence of tillite-like deposits (breccia-conglomerates), impoverishment of the strata in faunal remains, abundance of organic material in the rocks, and almost complete absence of carbonate interlayers permit assuming the presence of a cold moist climate at the time of accumulation of the terrigenous deposits. The South Verkhoyansk synclinorium is separated on the west along a

*We consider the terms *epigenesis* and *catagenesis* equivalent for postdiagenetic transformations of sedimentary rocks.

TABLE 1. Zones of Postdepositional Changes in Rocks of the Vilyui Syncline and VFZ

Vilyui syncline and western Verkhoyansk region [4]	Northern Verkhoyansk region, Kharaulakh, O. V. Yapaskurt, 1987	Southern Verkhoyansk region [1, 6]
Initial epigenesis — unaltered argillaceous cement	—	—
Deep-seated epigenesis — altered argillaceous cement and initial quartz—mosaic textures of sandstones	Deep-seated catagenesis — conformal regeneration textures	Deep-seated epigenesis and initial metagenesis — conformal regeneration structures, hydro-mica—chlorite cement and weak development of cleavage
Metagenesis — complex quartzite-like textures and sericite—chlorite cement	Initial metagenesis — blastic textures at contacts of quartz grains	Metagenesis (sericite—chlorite grade) — initial blastic deformation at contacts of quartz grains, developed cleavage
Metagenesis — spiniform textures and muscovite cement (greenschist facies)	Metagenesis — differential sliding structures and spiniform formations	
	Initial metamorphism — massive blastic, spiniform textures and segregation structures with metamorphic muscovite. Metamorphic biotite — in metasandstones and blastopsammitic quartz—mica schists	Metamorphism (biotite grade) — blastopsammitic and lepidoblastic textures. Metamorphism (staurolite grade) — epidote—amphibolite facies

series of faults from the Riphean—Lower Paleozoic terrigenous—carbonate formations of the Sette-Daban Range. In this structure participate rocks of the Verkhoyansk terrigenous complex from the Carboniferous to the Lower Triassic, crushed into linear folds of a northeast trend complicated by small folding of the second stage spatially related to zones of influence of deep faults. These zones, distinguished by a number of signs (isoclinal folding, intense cleavage, clustering of fractures), are traced within the investigated territory from the northwest to the southeast and control the location of zones of postdepositional transformation. Also widespread are open fractures, many of which are accompanied by copious hydrothermal manifestations and are ore-controlling.

Magmatism is manifested intensely in the investigated territory. Here are known granitoid massifs, the largest of which are the Uémlyakh and Tarbaganakh. The massifs have a multiple-phase structure. Also developed in the region are numerous dikes of lamprophyres and diorite-porphyrates, for which a preintrusive age was proved. A prebatholith origin of the greater part of the hydrothermal manifestations and, in particular, gold mineralization was also established.

The Verkhoyansk complex within the region has the following structure in general features.

Middle Carboniferous. Ékachan Suite. It is divided into three subsuites. The lower subsuite is represented by coarse-grained calcareous siltstones and fine-grained siltstones. Middle Carboniferous brachiopods and pelecypods are present. The middle subsuite is composed of black, finely stippled, bedded or massive siltstones. Thickness 250-300 m. The upper subsuite is composed mainly of fine-grained siltstones with infrequent concretions of a siliceous—siltstone and siliceous—carbonate composition. Thickness 250-350 m. Total thickness of the suite 900-1100 m.

Upper Carboniferous. Surkechan Suite. It is represented by coarse-bedded, coarse-grained siltstones with infrequent interlayers of fine-grained horizontal-bedded sandstones. Carboniferous productids are noted. Thickness 200-250 m.

Lower Permian. Khalyin Suite. The lower boundary of the suite is drawn along the base of coarse-grained siltstones, in which Lower Permian productids are found. Fine-grained sandstones with thin interlayers of gravelstones dominate in the lower part of the section. Up the section are mainly siltstones, black and dark-gray with individual sandstone members. At the top of the suite is a bed of gray fine- and medium-grained sandstones (up to 50 m). Total thickness of the suite 650-750 m.

Bonsalchan Suite. Black sandstones with individual beds of gray fine- and medium-grained sandstones dominate in the section of the suite. In the middle part is a bed of sandstones with a thickness up to 50 m. In the sandstones there is an abundance of flattened fragments of clay shales. Lenses of gravelstones and conglomerates with well-rounded pebbles are found. Thickness up to 1200 m.

Upper Permian. The total thickness of the Upper Permian deposits is about 3 km. It is divided into three suites.

Menkechen Suite. At the bottom are inequigranular polymictic sandstones, sometimes cross-bedded, with a subordinate number of interlayers of black clay shales. Higher up the section is an alternation of interlayers of black clay and sand-clay shales with interlayers of strongly calcareous sandstones. Thickness of the suite 1100-1300 m.

Chamba Suite. Is composed of massive gray fine- and medium-grained sandstones with subordinate interlayers of black clay shales, siltstone, and small-pebble conglomerates. Thickness of the suite 1000-1200

Imtachan Suite. The site is composed of fine- and medium-grained sandstones alternating with beds of clay, carbonaceous, and sand-clay shales. The section is topped by a bed (15-20 m) of light-gray sandstones with lenses and interlayers of small-pebble conglomerates. Thickness of the suite 600-900 m.

The **Lower Triassic** in the territory of the investigated region is developed limitedly and is represented mainly by medium- and coarse-grained light-gray sandstones and gravelstones.

PETROGRAPHIC CHARACTERISTICS

As was already mentioned, the cyclically constructed sections at different stratigraphic levels of the Verkhoyansk terrigenous complex are composed mainly of sandstones, siltstones, and clay shales.

In relatively weakly transformed varieties the polymictic sandstones are massive, more rarely vaguely horizontal-bedded, gray. The main rock-forming components are rock fragments, feldspars (albite, more rarely potassium feldspar), and quartz, single deformed and altered scales of muscovite and biotite. Sorting of the clastic grains is usually poor, roundness is also low, which indicates single-cyclicity and rapid accumulation of the sediments. Allogenic accessory minerals are represented by zircon (it dominates), tourmaline, garnet, ilmenite, titanomagnetite, hornblende, and pyroxene. An insignificant amount of clay material in the form of film and pore-film cement is present.

Among the rock fragments can be distinguished several groups: siliceous and siliceous-micaceous, acid extrusives and microgranites, albitophyres, clay shales and siltstones, and extrusives of an intermediate and basic composition. Even in weakly transformed varieties cleavage is noted in the form of uneven branching surfaces traced by inclusions of black organic material, the content of which in the rocks is on average 0.2%*.

Siltstones are present in members alternating with dark-gray clay shales. Their color is gray, the structures, as a rule are massive or vaguely horizontal-bedded, cleaved in various degrees, platy. They are close to sandstones in composition of the clastic components. The content of clay cement is small (up to 5%), carbonaceous material is present in a noticeable (up to 2%) amount. The latter together with clay minerals cements the clastic grains, as a result of which the rock texture has a complex reticulate appearance.

Silty clay shales, dark-gray, sometimes almost black due to the abundance (up to 3%) of organic material, often closely foliated as a result of cleavage. The groundmass of the rock is represented by a hydromica-chlorite matrix in a complex mixture with carbonaceous material. Terrigenous grains of silt size amount to 10-30%.

The evolution of the composition of the sandstones with respect to the stratigraphic scale can be judged from the classification triangular diagram (Fig. 1) constructed by V. D. Shutov's method [7]. The majority of symbols fell into the field of quartz-feldspathic graywacke and a part into the field of feldspathic-quartz graywacke and graywacke arkose. The Lower Triassic sandstone is characterized by a high quartz content. Plagioclase markedly dominates in the composition of the Upper Carboniferous and Lower Permian feldspars; potassium feldspars occur in Upper Permian deposits. The composition of the fragments changes at this boundary: whereas an acid association (siliceous and acid extrusives, albitophyres) dominate in the Carboniferous and Lower Permian sandstones, fragments of intermediate and basic extrusives occur in a noticeable amount in the Upper Permian sandstones. The role of chlorite in the cement of these rocks increases accordingly.

ZONES OF POSTDEPOSITIONAL TRANSFORMATION

There are objective difficulties in working out a unified system of gradations of postdepositional transformation from epigenesis to comparatively high grades of metamorphism. They are related to the fact that facies (or zones) of epigenesis and metagenesis do not correspond in their content to metamorphic facies [2], since mineral equilibria, on which

*A description of secondary structures and clay and authigenic minerals is given in the next section.

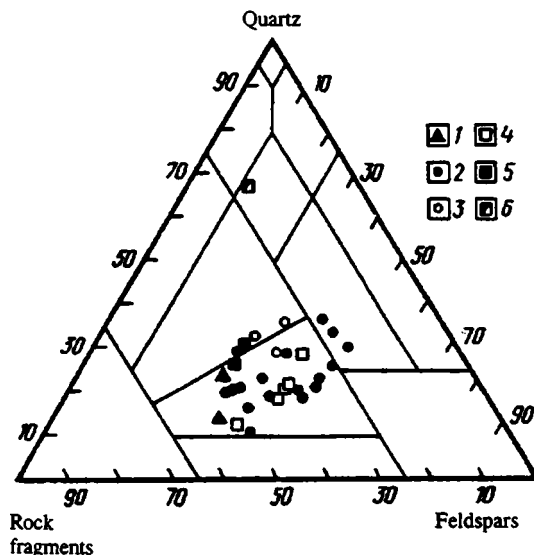


Fig. 1. Triangular diagram of arenaceous rocks of the South Verkhoyansk synclinorium: 1) Middle and Upper Carboniferous; 2) Lower Permian; 3) Lower Permian, metasandstones; 4) Upper Permian; 5) Lower Triassic; 6) Middle Triassic.

is based the principle of distinguishing metamorphic facies, are not established in relatively weakly transformed rocks. Therefore, when distinguishing gradations of postdepositional alterations the neutral term grade and concept zone (in the area and in the section) corresponding to it are used. A set of signs underlies distinguishing grades (zones) of transformation owing to the limitedness of the mineralogical data: mineral parageneses, character of transformation of terrigenous components, postdepositional structures of the rocks, cleavage.

Within the investigated part of the South Verkhoyansk synclinorium (Fig. 2) we distinguished the following zones (grades) of postdepositional transformation:

1. Deep-seated epigenesis and initial metagenesis.
2. Metagenesis (sericite-chlorite subfacies).
3. Metamorphism:
 - a) biotite grade; Greenschist facies
 - b) staurolite grade (epidote-amphibolite facies)

Deep-Seated Epigenesis and Initial Metagenesis. The changes in the arenaceous rocks of this zone are determined by the pressure-solution structure of the clastic grains and by authigenic minerals: hydromica, chlorite, pyrite, anatase, tourmaline, albite, and epidote.

In addition to the usual pressure-solution structures for epigenesis, characteristic structures caused by cleavage are distinguished. Cleavage cracks (when observed in a thin section) in general are subparallel among themselves but extremely sinuous and not persistent. Clastic grains contacting through the cleavage surface, as a rule, are mutually dissolved: their contacts have the form of typical pressure-solution structures. Consequently, during development of cleavage a part of the rock material passed into solution. The cleavage surfaces are seen distinctly, since they are filled with a nontransparent black material (amorphous iron and titanium oxides, as well as organics). Structures of differential sliding between clastic grains are also noted.

Clastic quartz at this stage of transformation is not altered and preserves the inherited typomorphic signs of a clastic mineral.

Feldspar, as a rule, is represented by albite, fresh or sericitized. The albite grains are often regenerated. Secondary segregations of albite xenomorphic with respect to the clastic grains, sometimes filling the pore spaces, are also noted. The size of the segregations of the newly formed albite reaches 1 mm.

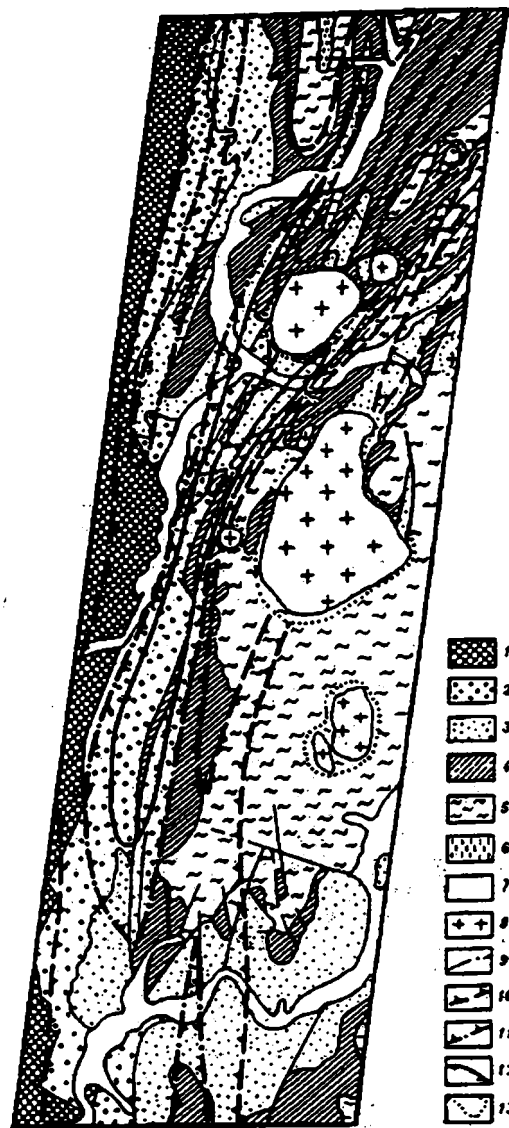


Fig. 2. Schematic diagram of metamorphic zonation of South Verkhoyansk synclinorium: 1) terrigenous-carbonate complex of Sette-Daban Range; 2) Middle and Upper Carboniferous; 3-4) Lower Permian (3 - Khalyin suite, 4 - Bonsalchan suite); 5) Upper Permian; 6) Lower Triassic; 7) Quaternary deposits; 8) granitoid massifs; 9) faults; 10) boundary of zones of epigenesis and metagenesis; 11-12) isograds (11 - of biotite, 12 - of staurolite); 13) zones of contact alteration.

Among the authigenic minerals of the titanium group, the most characteristic is rutile present in the form of prismatic crystals measuring 0.002-0.01 mm and forming complex sagenitic intergrowths (most often on albitophyre fragments). Rutile accumulations are often noted in the interstices between grains. Accumulations of partially crystallized (to anatase) leukoxene with a size of 0.5-0.1 mm are also present.

Apatite and tourmaline segregations are closely associated with albitophyre fragments.

Xenomorphous epidote segregations of an irregular palmate shape with a bright anomalous interference color are sometimes noted in the pore spaces between grains.

Hydromica of polytype $2M_1$ in a mixture with chlorite in various proportions is developed in the pore spaces and in the form of film cement. Clay cement, as a rule, amounts to not more than 5% of the rock volume.

The structures of clay shales are determined mainly by cleavage. Cleavage cracks, often outlined by organic material (with small pyrite segregations), are densely arranged and envelop the clastic particles; the structures as a result of this have a unique "micro-augen" appearance.

Relatively large, of the order of tenths of a millimeter, pyrite crystals are most characteristic for layers and lenses enriched with silt and fine-sand material but containing a considerable amount of organic material.

Thus, we refer to the zone (grade) of deep-seated epigenesis and initial metagenesis rocks whose structures and mineral parageneses were formed by processes of deep-seated epigenesis and were only insignificantly modified by superposed processes of initial metagenesis genetically related to processes of folding and metamorphism.

Metagenesis (sericite-chlorite subfacies of the greenschist facies). The boundary with the above-described zone is drawn on the basis of the occurrence of intense cleavage both in shale and in sand varieties and also on the basis of the initial occurrences of metamorphism of clastic quartz and recrystallization-granulation blastic deformation of its grains. In the minerals figuring in the name of the subfacies no signs distinguishing this zone from the preceding one are detected. Possibly, a more detailed chemical and x-ray diffraction study of sericite and chlorite would make it possible to use them as index minerals. True muscovite occurs only in the zone of metamorphism.

In arenaceous rocks of the zone of metagenesis both primary-clastic and deep-seated epigenetic structures (pressure solution) are preserved rather well. Metagenetic structures formed under conditions of dynamic metamorphism. Cleavage, distinctly displayed in shales, is found in sandstones along series of branching surfaces traced by inclusions of organic material. The characteristic authigenic minerals of the zone of metagenesis are albite, chlorite, sericite, epidote, pyrite, tourmaline, and sphene. Albite is developed both as newly formed porphyroblastic segregations and as a result of replacement of more basic varieties of clastic plagioclases. It should also be noted that the rocks of the zone of metagenesis are very strongly saturated with authigenic pyrite, which can serve as an additional criterion for distinguishing it (pyrite in far smaller amounts is found in rocks of higher and lower metamorphic grades).

Clay minerals are represented by hydromica of the muscovite type ($2M_1$). And admixture of degraded hydromicas, possibly, with expanding sheets, is observed in certain specimens. Kaolinite is present in almost all specimens. Its quantity is presumably at the 5% level. The chlorite content (to judge from the x-ray diffraction patterns, essentially magnesian) is within 5-10%.

Thus, in addition to pyrite, transformations of clastic quartz are the most distinct mineral criterion for distinguishing the zone. Metamorphism of clastic quartz [6] is displayed in the most initial form: on the basis of statistical calculations of the distribution of typomorphic signs, almost no specific features distinguishing it from unmetamorphosed clastic quartz are detected. In quartz grains from rocks of the zone of metagenesis is noted an increased frequency of granulation mosaicism, which is genetically related to another phenomenon — the development of structures of recrystallization-granulation blastic deformation at its grain boundaries. In connection with the low content of clastic quartz in the rocks, mutual contacts of its grains are not often found, and therefore the very fact of their presence indicates relatively high rock transformation temperatures, since migration of the boundaries of quartz individuals is a thermally activated process.

Metamorphism. The isograd of the *biotite zone* is easily established both under field conditions and from a study of thin sections on the basis of the mass occurrence of newly formed biotite.

Rocks of zones of metamorphism substantially differ in appearance and textures from those described above. Thus, the sandstones look like gray schistose rocks, sometimes of an explicitly gneissoid appearance. They are coarse-platy with respect to schistosity, on the planes of which abundant mica segregations are often observed. The texture of the rocks is due to the production of schistosity and transformation of the clastic components as well as to the formation of authigenic minerals, including porphyroblastic segregations. Such structures can be described by the terms *blastopsammitic* or *lepidoblastopsammitic*. Ghosts of clastic grains are usually clearly distinguishable. Rock fragments were subjected to maximum recrystallization.

Clastic quartz in arenaceous rocks is deeply metamorphosed, which is expressed in the almost complete absence of mineral inclusions and of a mineral-forming medium and by weak and uniform presence of defects. Lenticules of quartz of an explicitly authigenic origin are observed, especially often behind large grains, in "shadow" zones.

Relatively small (up to 0.2-0.3 mm) plagioclase grains most probably have a clastic origin. A distinct twin structure is characteristic for them. Large (up to 1-2 mm) plagioclase segregations having a typical porphyroblastic appearance are noted. Their form is tabular but with unexpressed crystallographic faces. The polysynthetic twin structure in these grains is, so to speak, lubricated, and they are usually sericitized or epidotized. It corresponds in composition to oligoclase Nos. 25-29. A feebly expressed zonal structure is characteristic, in which case the peripheral zones are represented by albite, which was least subjected to secondary alterations.

The rock fragments are almost completely recrystallized into a quartz—feldspathic matrix.

TABLE 2. Interplanar Spacings (in Å) and Reflection Intensities of Biotite

Specimen 6308				Specimen 6302/2			
<i>d hkl</i>	<i>I</i>	<i>d hkl</i>	<i>I</i>	<i>d hkl</i>	<i>I</i>	<i>d hkl</i>	<i>I</i>
9,97	95	1,815	2	9,98	82	2,223	5
4,94	2	1,682	26	7,05	4	2,178	10
3,706	22	1,542	15	4,66	3	2,092	2
3,344	100	1,507	2	4,55	3	2,046	48
3,177	13	1,481	2	4,23	3	1,998	14
2,975	2	1,443	14	3,70	20	1,918	2
2,926	7	1,370	15	3,531	8	1,897	2
2,770	3	1,330	2	3,31	100	1,850	2
2,714	4	1,317	3	2,98	10	1,828	2
2,626	15	1,263	12	2,925	8	1,818	20
2,515	51	1,224	2	2,818	14	1,682	2
2,436	11	1,138	2	2,715	4	1,627	2
2,272	2	1,124	2	2,622	12	1,543	13
2,222	6	1,069	7	2,522	45	1,443	13
2,179	11	—	—	2,444	9	1,358	12
2,019	51	—	—	2,307	2	1,226	3
1,914	2	—	—	2,293	2	1,153	3
1,850	2	—	—	2,275	4	1,123	5

Biotite is present in almost all specimens (1.5-22 vol. %). The form of the biotite segregations is irregular platy, xenomorphic with respect to the clastic grains. Certain large (up to 1 mm) biotite segregations replace entire groups of clastic grains and form typical porphyroblasts. Biotite intensely exhibits pleochroism according to the scheme: Ng – brown, Np – pale yellowish–brown. Double refraction is high, extinction is parallel. The indices of refraction of biotite Nm – 1.625-1.630 in the biotite zone and considerably higher (Nm ~ 1.660) in the staurolite zone. The data of an x-ray diffraction study of biotite (from the biotite zone) permit judging the complete trioctahedral character of the mineral (Fe³⁺ [6] and Al³⁺ [6] are absent in the octahedra). This follows from the virtual absence of 002 reflections (*d* = 5.0 Å); this is also indicated by the high value of line 060 with *d* = 1.543 and 1.542 Å. The almost equal intensities of reflections 004 and 005 as well as the rather high value of reflection 060 (*d* = 1.543 Å) suggest approximately equal ratios of Fe²⁺ and Mg²⁺ in octahedra, i.e., a moderate iron content of the mineral (Table 2). The parameters of the unit cells of the investigated biotites are close:

Specimen 6302/2 *b*₀ = 9.258 Å; *c* ≈ *c*₀·sin β ≈ 10.029 Å.

Specimen 6308 *b*₀ = 9.252 Å; *c* ≈ *c*₀·sin β ≈ 9.95 Å.

The biotite flakes in the rocks are oriented either in the direction of schistosity or perpendicular to it. Some biotite segregations are partially replaced by pale-green weakly pleochroic chlorite with a low anomalous interference color.

The content of newly formed muscovite is from 5 to 15%. This is either individual elongated laminae up to 0.5 mm long or bundles of the thinnest laminae which are elongated along the schistosity of the rocks and envelop the clastic grains or porphyroblastic segregations.

Minerals of the *epidote*–*zoisite* series are found only in certain specimens (up to 1-2%). Epidote (zoisite) segregations are present as small (up to 0.03 mm) barrel-shape grains, usually with a low anomalous interference color. Xenomorphic aggregates of this mineral (up to 0.5 mm) are noted, and the core of such aggregates is characterized by a high anomalous interference color. Sphene is closely related with epidote. It is usually present as aggregates of the smallest (<0.01 mm) granules, semitransparent, with very high relief and birefringence.

Pyrrhotite is present in insignificant amounts in about half of the specimens studied. It forms either individual irregular poorly crystallized granules measuring 0.005-0.02 mm, or openwork irregular aggregates of these granules. In reflected light the color is dark bronze with reddish reflections, it is magnetic.

Apatite is present as single poorly crystallized prismatic grains measuring 0.02-0.05 mm. Birefringence is very low.

Porphyroblastic garnet segregations are very rare and confined to the staurolite zone. Micropoikilitic inclusions of clastic grains are characteristic. Pale pink in thin sections, pink in grindings. The indices of refraction are *N* ~ 1.800-1.802. Apparently, garnet is represented by almandite.

Single segregations of authigenic tourmaline are noted.

Lenticular and irregular segregations of large calcite crystals are present in some specimens.

Flyschlike Metasiltstones and Clay Shales. Strongly schistose dark-gray rocks with uneven slab or foliated parting. The abundance of oriented micas (muscovite) on the planes of schistosity imparts a silky luster to them. "Nodules" of porphyroblastic segregations measuring up to 2 mm are clearly noticeable on the planes of schistosity.

The rocks are composed of newly formed micas (muscovite and biotite), quartz, feldspars, and carbonaceous material. The porphyroblasts are represented by minerals of the rhabdophanite—monazite—bastnäsite series and, more rarely, staurolite (in the staurolite zone of metamorphism). Pyrite and pyrrhotite are present in a noticeable amount.

Clastic quartz and feldspar grains in rare cases are discerned completely distinctly (their size is 0.02-0.15 mm, the grains are very elongated along the schistosity as a result of intense solution). More often feldspars are absent, and quartz is represented by lenticular formations, for which it is impossible to discern the origin (primary clastic or authigenic). Most probably, these are complex combinations of metamorphosed flattened clastic grains and newly formed quartz that grew in the direction of schistosity.

Muscovite segregations are greatly elongated, distinctly oriented along the schistosity, and envelop large quartz grains and porphyroblasts. Muscovite is overfilled with finely divided organic material imparting a dark-gray color to the rocks.

Biotite is contained usually in insignificant (up to 5%) amounts and rarely associates with muscovite, it most often replaces the quartz—feldspar formations.

Porphyroblasts of the rhabdophanite—monazite—bastnäsite series are found in shales of various metamorphic grades, but are confined only to the uppermost part of the Upper Carboniferous. In intensely transformed rocks (biotite and staurolite zones) "shadow" cavities filled with newly formed quartz are usually present along the schistosity on both sides of the porphyroblastic segregations. The porphyroblasts are often turned at an angle to the direction of schistosity (rotation structures). These facts most probably indicate the premetamorphic formation of rare-earth rhabdophanite concretions transformed during the course of metamorphism to monazite, which in turn is partially or completely replaced by rare-earth carbonate. Consequently, we can assume a diagenetic origin of the concretions of hydrous phosphates of rare-earth elements, which developed only at a certain stratigraphic level and were subsequently transformed along the monazite—bastnäsite series. Thus, the described segregations are not typical metamorphic porphyroblasts of so-called Khotenschiefer.

Staurolite is usually represented by single tabular or elongated prismatic segregations measuring 0.1-0.4 mm and is confined to the zone of the most intense transformation of the rocks (staurolite zone). The staurolite grains are extended along the schistosity of the rocks. Pleochroism in golden-yellow hues is noted. Cruciform twins are present. Optically positive, positive elongated. Low birefringence (0.012), its noticeable dispersion is observed. Staurolite is overfilled with inclusions of carbonaceous material hindering its diagnosis. Unlike minerals of the monazite series, shadow cavities filled with metamorphosed quartz do not form, as a rule, around the porphyroblastic staurolite segregations. Evidently, the staurolite segregations, as, by the way, those of biotite, are "true-porphyroblastic" unlike the "pseudo-porphyroblastic" grains of rare-earth phosphates.

Pyrite and pyrrhotite are common minerals in phyllite-like schists. They most often form lenticular-elongated (up to 1.5 m) irregular segregations confined to quartz lenticles.

On the whole the texture of the schists can be characterized as blastoaleuritic — up to lepidoblastic porphyroblastic.

It is necessary to note also the alterations of the sedimentary and metasedimentary rocks in the exocontact aureoles of granitoids. The near-contact hornfels differ clearly from products of zonal transformation both in characteristic textures and in newly formed minerals (cordierite, andalusite, amphiboles). Contact metamorphism is superposed in various parts on rocks of all grades of zonal metamorphism known in the region (see Fig. 2), which unequivocally proves the pre-batholith origin of zonal (dynamothermal) metamorphism.

FACTORS AND STAGES OF POSTDEPOSITIONAL TRANSFORMATIONS

Unmistakable tectonic control of the distribution of zones of postdepositional transformation exists in the area (see Fig. 2). Zones of highest metamorphism are confined to zones of influence of concealed deep fault, of an NNE strike. In the southern part of the region, where the faults converge into a single bundle, the biotite and staurolite zones reach a maximum (20-25 km) width. These zones are intersecting with respect to the stratigraphic boundaries and folds.

It is impossible not to note elements of vertical zonation of the zones of postdepositional transformation. Upper Carboniferous–Lower Permian rocks are subjected to relatively deep alteration; Upper Permian and Triassic rocks are not taken in, at least in the investigated territory, by metamorphism of the biotite and staurolite grades. Evidently, the zones of postdepositional transformation as large geologic bodies have a complex structures, and their form is controlled by two principal factors: by the deep-seated character of transformation and by the main tectonic structures of the region – concealed deep faults (see Fig. 2).

To reconstruct the sequence of events of postdepositional transformation, it is necessary to examine the history of tectonic development of the region in which four stages are distinguished: main folding, porphyrite (small intrusions), metamorphic, and batholith. The formation of the main fold structures is related to the first stage, in which case the territory formed as a highland. A large number of interleaved quartz veins formed at that time. The next stage began with the occurrence of three fracture systems: two shear (submeridional and sublatitudinal) and tension (northwest). Along these fractures were introduced porphyrite intrusions of sublatitudinal and northwest trends, which intersected the quartz sheet veins and metamorphosed them. Dikes of granite-porphyries intruded somewhat later along submeridional fractures. The porphyrite stage ended with hydrothermal activity: quartz veins healed the shear and tension fractures.

The next stage was metamorphic. The second phase of folding occurred in zones of influence of longitudinal deep faults. Folding was intense isoclinal, often overturned. The axes of small folds of these zones usually trend at small angles to the fold structures of the first phase of folding. Subsheet quartz veins and dikes of porphyrites were crumpled into small folds, i.e., the second phase of folding occurred after the introduction of prebatholith intrusions.

Dynamothermal metamorphism is genetically related to the second phase of folding. Evidently, considerable lateral compression and high temperatures are the cause of metamorphism in zones of influence of deep faults. The maximum values of these parameters can be estimated from the presence in rocks of staurolite in association with muscovite and quartz, which permits comparing this zone with the low-temperature subfacies of the epidote–amphibolite facies [2]. Consequently, the temperatures of metamorphism of the staurolite zone exceeded 500°C; the pressure limits of stability of the subfacies are 3–6.5 kbar. Quartz veins filling cleavage cracks are related to the metamorphic stage.

The last, batholith, stage, as already mentioned, was manifested only in hornfelsization of the rocks in the narrow exocontact aureole of the granite intrusions.

Evidently, epigenetic transformations of the rocks are confined to the preinversion stage of development of the miogeosyncline and, according to the notions of A. G. Kossovskaya and V. D. Shutov [4], were controlled by the thickness of the overlying strata. It can be assumed that the initial metagenetic processes were determined by the first phase of folding, and dynamothermal metamorphism is related to the second phase. Such an interpretation calls for three separate stages of postdepositional transformation of the rocks of the region, not counting contact metamorphism. However, actually this simply was not so.

Thus, it is impossible to imagine the formation of thick quartz sheet veins (filling of cavities and metasomatic) related to the first phase of folding as a result of the redistribution of silica during initial metagenesis. Most probably, silica was removed into the zone of initial metagenesis from the zone of metamorphism that existed in the first phase of folding but did not reach the level of the present-day erosion surface.

It seems to us that the tectonic and metamorphic processes of the orogenic stage of development of the Verkhoyansk miogeosyncline occurred interrelatedly, gradually, and continuously. Concrete manifestations of these processes projected onto the present-day erosion surface, obviously, did not always reflect their true sequence. Thus, for example, the entry of considerable portions of silica from deep-seated zones and the intrusion of dikes of diorite-porphyrites could have outstripped the gradual rise of the thermal front, which was especially intense in zones of influence of deep faults. A rise of temperature in deep-seated zones, probably, was reflected in the late development of anatectic granitization, to which is related relatively later introduction of granite intrusions.

Thus, we are inclined to the notion of two stages of postdepositional transformation of terrigenous rocks of the Verkhoyansk complex. The first, preinversion, stage is deep-seated epigenesis. The second, which gradually developed during the orogenic stage of development of the geosyncline, superposed on the regional plan of deep-seated epigenesis the entire range of zones of postdepositional transformation – from initial metagenesis to the epidote–amphibolite facies of metamorphism, inclusively.

One of the authors of the article (N. G. Andriyanov) considers that dynamothermal metamorphism (biotite and staurolite zones) is local and confined only to zones of influence of deep faults. However, it can be assumed that metamorphism is regional, and zones of influence of deep faults were heat conductors, which was reflected in the configuration of

metamorphic zonation. Analyzing the zonation map (see Fig. 2), we can easily imagine that the zones of relatively high metamorphism subsided under the zones of metagenesis and epigenesis (analogously to the alternation of anticlinal and synclinal structures), but there are no data on how steeply these zones subsided.

Here is no direct proof of the existence of zones of more intense metamorphism at depth, under the level of the present-day erosion surface. If we assume the existence at depth of metamorphic rocks of the amphibolite facies (higher than the sillimanite isograd), we should assume also the possibility of anatexis, which is almost always interrelated with conditions of the amphibolite facies. N. L. Dobretsov et al. consider that "...for granite bodies with deep roots one should expect the most distinctly manifested conformality of their contacts and metamorphic isograds of a regional plan. The transported masses of granitic melts that lost such a connection with deep roots, conversely, cannot have a substantial effect on the general picture of the temperature distribution in metamorphic strata, since their influence is virtually limited to supplies of stored heat" [2, p. 111].

The granitoid intrusions in the investigated territory are, evidently, precisely such "transported masses, and their contact effect (in narrow aureoles) on the enclosing rocks of different zones of postdepositional alteration is the third stage of lithogenesis of terrigenous rocks of the South Verkhoyansk synclinorium.

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CRYOGENIC PROCESSES IN THE GRANULOMETRIC COMPOSITION OF SEDIMENTARY FORMATIONS

Yu. L. Yurov

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Using a new technique of data representation it is shown that processes of brittle breakage of rocks play a leading role in the formation of the bimodal granulometric distributions discovered in various sedimentary formations on the Kola peninsula, such as the primitive cryogenic weathering crust and moraine. Stages in the transformation of the original bimodal distribution into a unimodal distribution as a result of solifluction displacement of dispersed material are outlined. Comparisons made on the basis of a unified technique for the description of terrestrial and lunar deposits has made it possible to gain a better understanding of the nature of both types of deposits.

The length of the era of continental development and the climate changes that occurred during this period are responsible for the diversity of the sedimentary deposits found on the Kola peninsula, and the comparatively quiescent tectonic regime has helped to preserve these deposits. These circumstances have made it possible to study a diverse range of sedimentary formations over a comparatively small area and, by making comparisons between these formations, to arrive at interesting conclusions. Such investigations become especially important when new technical approaches are involved. In studying continental deposits the granulometric composition is often the only source of information about the most important processes involved in the formation of these deposits. Thus, in [9] a new technique of data representation was employed in order to describe the granulometric composition, and on the basis of this technique, it was shown that processes of brittle breakage of rocks play a leading role in the formation of the primitive cryogenic eluvium of alkali rocks. In the present article the technical approach discussed in [9] is extended by means of computers and applied to material from other sedimentary formations.

To avoid terminological imprecision, we will have to define a number of concepts employed in the description of the granulometric composition by means of probability theory [1]. The tables that are generally used to represent the results of granulometric analyses (where the dimensional ranges of the fractions are given in one column, and the proportion of the total mass either in percentages or in decimal fractions of a unit, are given in parallel columns, for the corresponding natural formations) at first glance recall a grouped statistical series. One difference is that in grouped statistical series it is statistical frequencies p_i^* which are specified, whereas in tables we are concerned with the proportions of the total mass. However, it may be shown that this difference is purely formal in character. Though the very process of granulometric analysis has little in common with a statistical testing scheme, the results obtained in the course of statistical tests may be related to corresponding abstract schemes. This is the case with results that are expressed both in terms of particle counts (where each individual particle is associated with a corresponding fraction) and in terms of units of mass (where an infinitely small mass is associated with the discharge corresponding to the particle from which this mass has been taken). The limits of the discharge, expressed in terms of the grain sizing of the fractions, may be given in any one of a number of forms, for example in terms of linear dimensions or in terms of the logarithms of the linear dimensions. The first method is more conveniently applied in text and the second in diagrams or calculations. That it is possible to pass from one dimension to another demonstrates that these dimensions are essentially the same, while the diversity of the logarithmic scales denoted by the Greek letters is redundant. In the opinion of the present writer, the scale based on natural logarithms is the most convenient and easiest to use. To make it easier to discern the information presented by means of these scales, wherever necessary, we can always specify not only the logarithms, but also the linear dimensions, and then limit the discussion to the latter alone. In the present article the logarithmic scale is denoted z , i.e., $z = \ln x$ (where x is the linear dimension of the particles, in mm). Table 1 illustrates the grouped statistical series we have referred to, though there are a number of differences from similar tables that have been used in granulometric analysis. To help in applying probabilistic methods the proportions of the total mass of the fractions are expressed in terms of parts of a unit, rather than in percentages. All the results of the analyses conformed to the requirement imposed on the precision of a sieve analysis,

TABLE 1. Granulometric Composition of the Primitive Cryogenic Weathering Crust of Gabbro

Fraction	Size Limits of Fractions, mm	Logarithm of size limits	Proportion of Total Mass of Fractions of Samples of Primitive Cryogenic Weathering Crust		
			Sample V 42.6 (depth 0.0-0.2 m)	Sample V 42.5 (depth 0.4-0.6 m)	Sample V 42.3 (depth 0.8-1.0 m)
	0,0002 *	-8,517			
1	0,002	-6,215	0,0240	0,0382	0,0431
2	0,01	-4,605	0,0829	0,0902	0,1051
3	0,05	-2,996	0,2348	0,2536	0,2493
4	0,063	-2,764	0,0195	0,0248	0,0207
5	0,10	-2,303	0,0416	0,0557	0,0435
6	0,16	-1,833	0,0315	0,0433	0,0348
7	0,20	-1,609	0,0161	0,0221	0,0178
8	0,315	-1,155	0,0365	0,0495	0,0416
9	0,40	-0,916	0,0196	0,0280	0,0225
10	0,63	-0,462	0,0379	0,0490	0,0412
11	1,00	0,0	0,0440	0,0543	0,0426
12	1,60	0,470	0,0738	0,0763	0,0643
13	2,50	0,916	0,0693	0,0651	0,0591
14	10,00	2,303	0,2695	0,1499	0,2144

*The minimal size of the particles is tentative.

i.e., the wastage of mass did not exceed 0.02 [4, p. 101]. This made it possible to represent the results in reduced form, i.e., in such a way that the sum of the proportions of the total mass of the fractions for each sample was precisely equal to one, as must be true in the sense of the dimension employed. Moreover, Table 1 as well as the succeeding tables give not the dimensional intervals, but rather the size limits, which are correspondingly shifted relative to the proportions of the total mass of the fractions (which is a more concise mode of expression, since the set of sieves is printed once, and is also more comprehensible).

A distribution polygon is a graphic illustration of a grouped statistical series. It is not important how the diagram is constructed, whether as a broken line or stepwise, or, finally, as a smoothened curve, but if it is constructed in a coordinate plane in which the percentages or proportion of the total mass of the fractions are laid out along the vertical axis, the result is a distribution polygon, and not a histogram. The use of this type of diagram to evaluate the distribution law of a random variable may — if the magnitude of the discharge is not kept constant — lead to errors that can have catastrophic effects [5, 19].

A rough idea of the law of a random variable may be obtained by means of a table of frequency densities f_i^* (Table 2). The values of the quantities that make up this type of table are found by dividing each statistical frequency, p_i^* , by the magnitude of the corresponding discharge $\Delta z = \ln x_{i+1} - \ln x_i$. The quantity that is obtained as a result is dimensionless, since the quantities from which it is obtained by division are dimensionless. Histograms in the form of a set of rectangles each of which characterizes a definite discharge or fraction (diagram), represent a graphical illustration of these types of tables. The area of one of the rectangles, when expressed in terms of the corresponding scale, is equal to the statistical frequency p^* of the discharge (in grain sizing, to the proportion of the total mass of the fraction), and its base is equal to the magnitude of the discharge (in grain sizing, to the discharge interval of the fraction). In this scheme for representing the results of granulometric analyses, both frequency density tables and the histograms are intended to serve mainly for determining the number of terms j in the equation

$$D = \sum a_j d_j,$$

TABLE 2. Frequency Density of Granulometric Composition of Primitive Cryogenic Weathering crust of Gabbro

Fraction	Size limits of Fractions, mm	Logarithm of Size Limits	Frequency Density of Samples of Primitive Cryogenic Weathering crust		
			Sample V 42.6	Sample V 42.5	Sample V 42.3
1	0,0002	2,302	0,010	0,017	0,019
2	0,002	1,610	0,051	0,056	0,065
3	0,01	1,609	0,146	0,158	0,155
4	0,05	0,232	0,084	0,107	0,089
5	0,063	0,461	0,090	0,121	0,094
6	0,10	0,470	0,067	0,092	0,074
7	0,16	0,224	0,067	0,098	0,079
8	0,20	0,454	0,080	0,109	0,092
9	0,315	0,239	0,082	0,117	0,094
10	0,40	0,454	0,083	0,108	0,091
11	0,63	0,462	0,095	0,118	0,092
12	1,00	0,470	0,157	0,162	0,137
13	1,60	0,446	0,155	0,146	0,133
14	2,50	1,387	0,194	0,108	0,155
	10,00				

where D is the total distribution function $F(x) = P\{X < x\}$; D_j , distribution function describing the individual terms and having the form $D_j = 1 - R_j$, moreover, $R_j = \exp(x/k_j)^{n_j}$ are Rosin-Ramler equations.

Determining the coefficients that occur in the equations is most efficiently done by bearing in mind the types of problems that are to be solved and their lithological trend. In this case a_j is a coefficient that describes the proportion of the overall distribution represented by a given constituent. Obviously, $\sum a_j = 1$; k_j is a modal dimension, since if $z = \ln k_j$, the derivative of this constituent, $D_j'(z)$, has a maximum; and n_j is the gradation factor, the greater the value of which, the steeper is D_j and the more compressed is $D_j'(z)$ relative to the z -axis.

The number of terms in the initial distribution corresponds to the number of maxima in the histograms or, what is the same thing, in the tables of the frequency densities. It follows from Fig. 1 and Table 2 that the data which they present characterize distributions that consist of two terms, since there are two maxima, or modes, separated by a minimum. The oscillations between the maxima, which are of nearly equal value, and the minimum attain the value of the latter. Deviations that are random in nature and are an order of magnitude smaller are superposed on this basic law.

Distinguishing the components of the original distributions by means of computers does not represent a very complicated procedure. For this purpose, a program was written in Basic. The author is not an expert in programming, so that the program will not be presented here, though it will be useful to consider the basic stages in the implementation of the program. Once the value a_1 has been selected arbitrarily, the proportions of the total mass of the fine fractions p_i^* , the sum of which is less than this quantity, is divided by a_1 . As a result the aggregate of the proportions of the total mass of the fractions of the fine constituent $p_{1,i}^*$ are found. The values of $p_{1,i}^*$ are determined by means of successive subtraction of the quantities $R_{1,i}^*$ from unity. Next, on the basis of the relation $\ln(\ln(1/R)) = n(\ln x - \ln k)$, which is obtained by taking the natural logarithms of the Rosin-Ramler equation, the coefficients n_1 and k_1 are calculated by means of the method of least squares using the pair of values $R_{1,i}^*$ and $z_i = \ln x_i$. To determine the coefficients of a more approximate term, we proceed in precisely the same fashion, but now on the basis of the proportions of the total mass of the fractions that were not used in calculating the first constituent.

Once all six coefficients have been determined, the theoretical distribution frequencies $p_i = R_i - R_{i+1}$ are calculated by means of substitution in the equation $R = R_1 + R_2$. Using the theoretical (p_i) and the initial (p_i^*) frequencies, the values of $\chi^2/N = \sum(p_i^* - p_i)^2/p_i$ are determined. From all the possible sets of coefficients, the set in which the

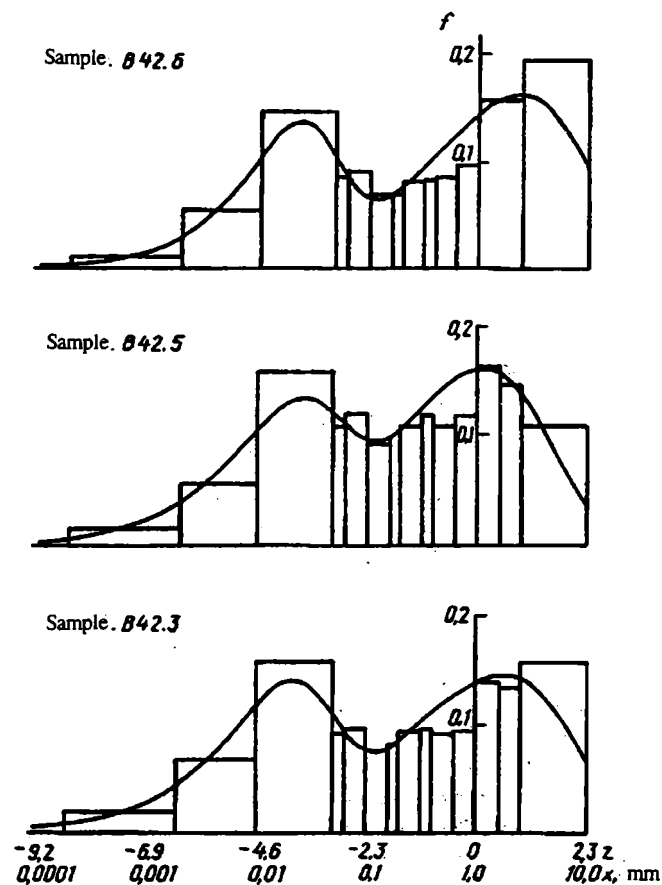


Fig. 1. Histograms and theoretical distribution density curves for the granulometric composition of the primitive cryogenic weathering crust of Gabbro.

quantity χ^2/N is smallest in value is discarded. The results of the calculations for samples from the primitive cryogenic weathering crust of gabbro are presented in Table 3. It was difficult to directly apply the Pearson criterion χ^2 to estimate the correlation between the theoretical equations and the experimentally determined data in the granulometric analysis was made difficult because of the fact that the number of tests N was not given. However, to make a selection between theoretical equations of the same type describing the same set of experimental data, we can make full use of the quantity $\chi^2/N = \sum (p_i^* - p_i)^2/p_i$.

We could also try to estimate the value of N in the more general case in which it is necessary to make a choice between theoretical equations with different numbers of components. Since the admissible error of the granulometric analysis is less than 0.02, and it is extraordinarily difficult to undertake tests with a lesser error, that, although hypothetical, are, in actuality, entirely feasible, it may be assumed that fewer than 50 tests will correspond to a single actual granulometric analysis. (If the number one is divided by a number greater than 0.02, a quantity less than 50 is obtained.) That is, the quantity χ^2 will be less than the product $(\chi^2/N)50$ and if the probability p corresponding to the quantity χ^2 is not small for a given number of degrees of freedom (cf. Appendix 3 in [1]), the probability corresponding to χ^2 will be even higher under these conditions and the divergence between the theoretical and the actual distributions may then be assumed to be insubstantial and is attributable to random causes. By comparison with the test of goodness of fit proposed by R. Schleyer [18], the use of the χ^2 criterion or $(\chi^2/N)50$ for investigating the granulometric composition is to be preferred, since the latter takes into account the entire aggregate of the differences between the experimental and the theoretical distributions.

TABLE 3. Coefficients of the Theoretical Equations of the Granulometric Composition of the Primitive Cryogenic Weathering Crust of Gabbro

Sample	a_1	k_1	n_1	a_2	k_2	n_2	n	χ^2/N
B42.6	0,2970	0,0218	1,036	0,7030	2,3850	0,623	0,746	0,0235
B42.5	0,3150	0,0211	0,869	0,6850	1,1452	0,638	0,711	0,0319
B42.3	0,3380	0,0185	0,896	0,6620	1,7613	0,591	0,694	0,0261

In order to verify the applicability of the method we are proposing for verifying the fit between the theoretical results and the experimental data, let us consider in more detail the granulometric composition of sample V42.6. Together with the proposed bimodal distribution (Table 3), the parameters of the Rosin–Ramler unimodal distribution may also be calculated for the granulometric composition. The values of the corresponding coefficients are, for the modal dimension k , 0.897, and for the gradation factor n , 0.515. For the theoretical unimodal distribution determined by these coefficients, the measures of the difference from the experimental data amount to $\chi^2/N = 0.367$ and $(\chi^2/N) = 18.350$. For r degrees of freedom, with r equal to $14 - 3 = 11$, the latter quantity corresponds to a probability lying in the interval from 0.05 (19.68) to 0.10 (17.28) (cf. Appendix 3 in [1]), so that there is no basis for claiming that there is a contradiction between the theoretical equation and the experimental data. At the same time, for a bimodal distribution $(\chi^2/N)50$. For a total number of degrees of freedom $r = 14 - 7 = 7$, we may associate with this quantity a probability greater than 0.90 (2.83) (cf. Appendix 3 in [1]). In this case the deviation of the theoretical distribution from the observed distribution may now be ascribed to random factors, and of the two different distributions – unimodal and bimodal – we are forced to choose the latter.

This method of representing the granulometric composition of sedimentary formations allows us to determine the coefficient n , which is equal to $\Sigma a_j n_j$ and which characterizes the overall gradation of the composition. The proposed coefficient, which takes into account possible differences in the number of modes of the granulometric distributions, yields an objective correlation on the basis of this characteristic of the lithological deposits. By comparison with the analogous quantities enumerated in a highly detailed list in [13] and commonly used at the present time, the special advantage of the proposed coefficient is that it is based on the entire smoothened aggregate of the experimental data. In the present study the computed values of the coefficient are presented in tables.

In the article mentioned earlier [9], a method of distinguishing the constituents of the granulometric composition of sedimentary formations was implemented for the case of the deposits in the sedimentary mantle of Khibina. The author is not able at present to solve this problem by means of computers, and the coefficients presented here characterize the theoretical distributions. With regard to these distributions there is no basis for claiming that their difference from the experimental data in all possible variants is minimal. By means of an exhaustive search through all the variants on a computer and comparisons between the variants and the quantity χ^2/N this part of the general problem has now been successfully solved, and the results are presented in Table 4. Because of the circumstances mentioned earlier, they differ somewhat from those presented in [9], though these differences fit entirely within the proposed scheme in which the granulometric compositions was formed.

Through the application of computers it has become possible to consider previously examined data from a new approach and to discover additional features that had not been previously noted. Consider the difference between the theoretical distributions of the near-surface samples of cryogenic residual rock. Material from Mount Norkpaht is significantly coarser than from the Rasvumchorr plateau. The modal coefficient of the term describing the finer part of the granulometric composition is found to be in the range 0.06 to 0.11 in the former case, and in the range 0.03 to 0.05 in the latter case. These figures are most likely due to the difference in the mineral composition of the original rocks. At point A1 the original rocks are coarse-grained khibinite enriched with unstable nepheline. The feature we have identified is manifest especially clearly when the results of the near-surface samples are compared: they are 0.11 and 0.03. However, in this case superposition of cryogenic sorting of the residual rock may also occur, leading to the formation of winterkilled patches, medalions, and analogous near-surface formations. One important feature which has been noted previously and has been confirmed in the present study is vertical differentiation with respect to the granulometric composition of the profile of the primitive cryogenic weathering crust along with the existence of three subhorizontal zones, an upper, middle, and lower zone. Of the three zones, the greatest quantity of fine particles is typical of the middle zone. This type of differentia-

tion is obvious even in field investigations when making comparisons between the quantity of silt and coarse-fragmental material. It is apparent that the results of analytic investigations of the silt confirm the field observations.

The use of an improved gradation factor shows that within the confines of a primitive cryogenic weathering crust gradation is most typical of the near-surface zones, where this coefficient may reach a value of 1.911. This may be attributed to inwash of fine weathering products into the deposit mass. However, for this process to attain its most pronounced level of development, the original conditions must be prescribed somewhat. Not only some difference in the resistance to weathering of various minerals in the original rock, but also their quantitative relationship and the rather high content of unstable minerals all appear to play a role here. By taking into account these elements it also becomes possible to understand why a maximum value of the gradation factor is typical of the sample from the near-surface zone of the primitive cryogenic weathering zone of iolite, whereas a minimum value is typical of samples from the weathering crust of gabbro.

Conducting the calculations on computers made it possible to reveal an additional feature of the granulometric composition of the solifluction deposits. From Table 4 it is clear that two bimodal theoretical curves for each sample were determined from point A2, and two sets of coefficients (one set given in parentheses) are correspondingly presented. The sets presented in parentheses are characterized by lesser proportions of the fine constituent, a lower value of the modal dimension and a higher value of the gradation factors and χ^2/N . The high value of the latter forces us to assume that the sets of coefficients that are not presented in the table within parentheses provide a better description of the corresponding experimental data. At the same time, since these coefficients are not so large, once we know the position of a solifluction deposit in the overall lithogenesis scheme, we may then assume that the distributions presented within parentheses refer to the original product of brittle breakage of rock which has subsequently undergone extensive transformations as a result of viscoplastic shifting of a stone river. The fact that the modal coefficients presented within the parentheses are close to the values of the same coefficients for the cryogenic residual rock confirms this hypothesis.

The data on solifluction deposits that have been presented are arranged in the form of a regular series that is in close agreement with the processes involved in the formation of the deposits. Initially, the products of brittle breakage of rock lose their finest particles in the course of solifluction shifting. As a result the gradation of the fine constituent grows, though the modal dimension remains within the initial range (sample A6). Further discharge of fine particles as a result of the development of the solifluction process leads to an increase in the modal dimension of the fine constituent, though it is still possible to discover traces of the original distribution (sample A2). Finally, there occurs a radical change in the original bimodal distribution of the products of brittle breakage of rock which turns the distribution into a unimodal distribution that is already a reflection of a viscoelastic process of a solifluction mud stream (sample A8.1).

Comparison of the gradation of the primitive cryogenic weathering crust and of the solifluction deposits shows that the material forming the latter is, on the whole, more highly graded. Thus, the entire mass of the stone river is characterized by a generalized gradation factor greater than 1.5. Liqueescency and coursing of the material reduces its gradation, bringing the value of the generalized coefficient down to 0.711.

The granulometric composition of the primitive cryogenic weathering crust of gabbro sets it apart from the primitive cryogenic weathering crust of the eluvium of alkali rock not only by virtue of the poorer degree of gradation, but also by virtue of the smaller value of the modal dimension of the finer constituent. This differentiation may also be attributed to the difference in the mineral composition of the original rock, in that the fine constituent of the eluvium of gabbro is determined by the products of the breakage of plagioclase rocks, whereas the major proportion of the eluvium of alkali rock consists of K-Na feldspar. According to the data presented in Table 3, the greatest quantity of fine particles and the lowest values of the modal and gradation factors are natural to the sample taken from the interface between the deposits and the original rocks, and not from the central part of the profile of the weathering crust. However, this does not contradict the law that has been established for the existence of three zones and the fact that the maximum degree of expression of the weathering processes is found in the middle of the three zones, and only confirms the field observations of the subhorizontal nature of the development of these zones, which is complicated by numerous streaks, "pockets," and other manifestations of cryogenesis.

If we wish to compare the granulometric composition of sedimentary formations that have developed under the conditions of a cold moist climate with deposits associated with glacial lithogenesis, we could consider the granulometric composition of moraine. The initial results of the granulometric analyses from which the data of Tables 5 and 6 were obtained are borrowed from [6]. The fact that the granulometric composition of moraine is inherent to the bimodal

TABLE 4. Coefficients of Theoretical Equations of the Granulometric Composition of the Sedimentary Formations of Khibina

Sample	a_1	k_1	n_1	a_2	k_2	n_2	n	χ^2/N
A1.1	0,7100	0,1103	0,504	0,2900	3,3096	1,867	0,900	0,027
A1.2	0,8150	0,0758	0,524	0,1850	3,3621	1,975	0,793	0,017
A1.3	0,4840	0,0594	0,641	0,5160	4,2379	1,490	1,079	0,017
A5.3	0,0830	0,0311	0,607	0,9170	4,6740	2,029	1,911	0,002
A5.4	0,5280	0,0301	0,675	0,4720	2,2982	1,421	1,027	0,002
A5.5	0,7980	0,0434	0,700	0,2020	3,3483	1,528	0,868	0,048
A5.6	0,6070	0,0460	0,611	0,3930	2,8949	1,707	1,042	0,037
A2.2	0,3900	0,4462	0,873	0,6100	3,5585	2,008	1,565	0,021
(A2.2	0,0670	0,0338	0,989	0,9330	2,4322	1,013	1,013	0,038)
A2.3	0,3700	0,3956	0,813	0,6300	4,0758	1,943	1,525	0,024
(A2.3	0,0770	0,0339	0,906	0,9230	2,9302	0,941	0,939	0,043)
A2.4	0,4400	0,4128	0,935	0,5600	3,5696	2,031	1,549	0,032
(A2.4	0,0830	0,0386	1,111	0,9170	2,2656	0,971	0,983	0,040)
A6.1	0,3380	0,0401	1,186	0,6620	1,9505	1,150	1,163	0,019
A6.2	0,6730	0,0512	1,185	0,3270	2,1860	1,231	1,201	0,064
A6.3	0,3120	0,0414	0,987	0,6880	3,2669	1,080	1,051	0,034
A8.2	0,2250	0,0290	0,954	0,7750	1,7528	1,030	1,013	0,031
A8.3	0,2800	0,0349	1,104	0,7200	1,4305	0,956	0,998	0,042
A8.1	1,0000	0,4791	0,711	—	—	—	0,711	0,072

Remark. A1 (1, 2, 3), samples of residual rock from the peak of Mount Norkpaht (the numbers 1, 2, and 3 refer to the upper, middle, and lower zones, respectively, of the weathering crust); A5 (3, 4, 5, 6), samples of residual rock from the peak of the Rasvumchorr plateau, with 3 representing material from the surface, and 4, 5, and 6 the upper, middle, and lower zones, respectively, of the weathering crust; A2, ... , A6, ... , and A8, samples of solifluction deposits; A2, southern slope of Mount Norkpaht; A8, the southwest slope of Mount Norkpaht; A6, northern slope of Mount Lyavochorr; A2.2, material lying directly beneath blocks from a stone stream; A2.3, A6.2, and A8.2, samples of the middle zones of the solifluction deposits; A2.4, A6.3, and A8.3, samples from the interface with the subjacent rocks; A6.2, material from the surface of a solifluction terrace; A8.1, material from a solifluction mud stream.

distribution of the particles has long been known [14], and the data presented by S. A. Strelkov and B. I. Koshechkin confirm that the moraines of the Kola peninsula do not represent an exception to the rule. However, through the application of a new technique in the analysis it has become possible to compare a moraine, in terms of this property, both with itself and also with other sedimentary formations.

Note the trend towards coarsening of the finer constituent (k_1 increasing from 0.09 to 0.12) as well as the growth of the proportion of this constituent (a_1 increasing from 0.77 to 0.86), both of which clearly seem to grow with depth. The assumption of subaerial diagenesis and the predominantly physicochemical breakage under the near-surface conditions of the original moraine material, followed by inflow of fine particles into the lower-lying stratum, is in good agreement with this law. The possibility for inflow of finer particles is also attested to by the reduction in the overall gradation factor n , from 0.93 (in the upper zone) to 0.89 (in the lowest zone). The same trend for this probable process is exhibited by the gradation factor of the fine constituent n_1 . It is clear from the values of the variables we have presented that the processes of subaerial diagenesis are not very highly pronounced, though by resorting to mathematical methods, we find we have a basis for saying that these processes do indeed exist.

Comparison of the granulometric composition of moraines with primitive cryogenic weathering crusts reveals a feature that the two materials share in common, i.e., their distributions are both bimodal. The resemblance of the distributions may be seen not only in general outline, but also, conspicuously, in the frequencies, in each of the groups of fractions that were identified. This is because the physical nature of the dominant processes underlying their formation are the same and the fact that, in both cases, processes of brittle breakage of rock play the principal role. However, there are also differences, which are determined by the greater degree of expressiveness of the accompanying processes in the formation of the primitive cryogenic weathering crust. The fine constituent of moraine exceeds, both quantitatively (a_1 in the range 0.77 to 0.86) as well as in terms of modal dimension (k_1 in the range 0.09-0.12), the analogous indicators of cryogenic eluvium (a_1 in the range 0.08-0.82 and k_1 in the range 0.02-0.11). Note, too, the greater scatter of the parameters that are associated with a primitive cryogenic weathering crust. This feature may be explained by noting that the deposit we are considering is currently undergoing a stage of intensive development, hence it is natural for the intensity of the processes forming it to exhibit some variability.

TABLE 5. Frequency Density of Granulometric Composition of Moraine

Fraction	Size Limits of Fractions, mm	Limiting intervals (logarithms)	Frequency Density of Moraine Samples*		
			sample 75-63/4 (depth 2.5 m)	sample 75-63/5 (depth 3.9 m)	sample 75-63/6 (depth 5.2 m)
1	0,0002	-8,517	0,020	0,033	0,026
2	0,002	-6,215	0,060	0,053	0,043
3	0,01	-4,605	0,141	0,141	0,140
4	0,05	-2,996	0,281	0,266	0,273
5	0,063	-2,764	0,206	0,196	0,225
6	0,10	-2,303	0,163	0,149	0,189
7	0,16	-1,833	0,140	0,140	0,178
8	0,315	-1,155	0,120	0,120	0,160
9	0,40	-0,916	0,082	0,086	0,097
10	0,63	-0,462	0,075	0,081	0,078
11	1,00	0,0	0,072	0,085	0,056
12	1,60	0,470	0,081	0,098	0,061
13	2,50	0,916	0,093	0,077	0,070
	10,00	2,303			

*The samples were taken from the wall of a quarry situated 1 km to the north of Lake Travyanyi.

TABLE 6. Coefficients of Theoretical Equations of Granulometric Composition of Moraine

Sample	a_1	k_1	n_1	a_2	k_2	n_2	n	χ^2/N
75-63/4	0,7740	0,0877	0,735	0,2260	3,4581	1,611	0,933	0,0210
75-63/5	0,8145	0,0952	0,626	0,1855	3,2706	2,167	0,912	0,0253
75-63/6	0,8560	0,1150	0,678	0,1440	3,8209	2,177	0,894	0,0215

The fact that the granulometric composition of moraine resembles a primitive cryogenic weathering crusts, a result we have been led to from a superficial treatment, may be interpreted as a further argument in favor of suggestions that the two should be examined in conjunction [2]. That these formations are genetically distinct was demonstrated exhaustively in [10]. Bearing in mind that both distributions of the particles by dimension are bimodal, the fact that the formations are genetically distinct may be taken as an additional proof that here we have a different type of deposit the gradation of which is also an intrinsic feature of the distinction, but for which the possibility of glacial genesis is excluded. Tables 7 and 8, which contain data on lunar regolith, illustrate these points.

If we compare the granulometric composition of all three types of sedimentary formations (primitive cryogenic weathering crust, moraine, and lunar regolith), we are led to conclude that they are extraordinarily similar, both in terms of the overall nature of the distributions, as well as in terms of frequencies, i.e., the constituents of the distributions. If we next compare the conditions under which each of these natural deposits developed, we are led to attribute the similarity of the compositions to the existence of a physical process that is common to all the deposits. On the basis of general assumptions of the physics of the grinding process [8], we are compelled to recognize brittle breakage of primitive rocks and the minerals composing these rocks as such a process. The parameters of the fine constituent (which in fact also determines the overall nature of the distributions) would appear to be related to the magnitude of the Griffith critical fracture. Since the interatomic distances and size of the reduction in strength as a result of the development of microscopic fractures are relatively close in many minerals, the ranges of these quantities in different minerals are not very different. According to

TABLE 7. Frequency Density of Granulometric Composition of Lunar Regolith

Fraction	Size Limits of Fractions, mm	Frequency Density		Size Limits of Fractions, mm	Frequency Density (sample 71061.1)
		Sample 60009.456	Sample 60009.457		
1	0,002	0,075	0,049	0,002	0,078
2	0,010	0,168	0,116	0,020	0,151
3	0,020	0,181	0,155	0,045	0,165
4	0,045	0,187	0,210	0,075	0,165
5	0,075	0,179	0,180	0,090	0,169
6	0,090	0,163	0,187	0,150	0,138
7	0,150	0,150	0,174	0,250	0,102
8	0,250	0,124	0,158	0,500	0,050
9	0,500	0,085	0,108	1,000	0,088
10	1,000	0,065	0,068	2,000	0,097
11	2,000	0,066	0,134	4,000	0,111
12	4,000	0,101	0,072	10,000	0,066
	10,000			40,000	

Remark. Data on samples 60009.456 and 60009.457 are taken from [17]; data on sample 71061.1 are taken from [15].

TABLE 8. Coefficients of Theoretical Equations of Granulometric Composition of Lunar Regolith

Sample	a_1	k_1	n_1	a_2	k_2	n_2	n	χ^2/N
60009.456	0,7800	0,0881	0,744	0,2200	4,3180	1,125	0,828	0,0449
60009.457	0,7670	0,1287	0,806	0,2330	3,5059	1,680	1,010	0,0279
71061/1	0,6070	0,0727	0,820	0,3930	6,0550	0,902	0,852	0,0148

the results of one study [8], in quartz the Griffith critical fracture is in the range 1.5-600 m. The fit to the modal dimension of the thin constituent of the samples under study is highly eloquent. In view of the physical nature of these samples, the groups of fractions that have been identified in the composition of these deposits are seen to fall, somewhat provisionally, in a range from the thinnest, which we take to be a constituent of the mineral fragments, to the coarsest, which we take to be a constituent of the mineral grains.

Drawing comparisons between the grade analysis of terrestrial and lunar samples is not a truly abstract pursuit. The conclusions that may be drawn on the basis of a grade analysis represent a contribution to our knowledge of both types of samples. With regard to terrestrial deposits, it follows from such a comparison that the bimodal nature of a granulometric composition does not in and of itself provide a basis for classifying sedimentary deposits as moraines. At the same time, the conclusion therefore follows that a bimodal distribution is far from always a necessary consequence of the mixing of mature and immature regolith [16].

An understanding that material which has been subjected to a grinding process will inevitably exhibit a bimodal distribution comes with the force of an immediate experience to anyone who has ever tried to shatter rocks in the mortar of a pestle. Without repeated sifting it is virtually impossible to crush a sample if its volume exceeds some value at which it is possible to distribute, or "smear," it in the form of very fine layer on the walls of the mortar. Where the fine and coarse particles are in a definite ratio, particles of the former type, because they resist brittle breakage, will form an aggregate mass that absorbs the basic loads and, as a consequence, the overall rate of the grinding process drops sharply. One way out of this situation is to sift the sample and then shatter only the material that is left on the sieve. In these circumstances sedimentation under natural conditions will be unable to "sift" material that has been subjected to grinding no matter how much time there is available for realizing such a rock breakage process. It is, thus, not remarkable that among the samples of lunar regolith there are those in which the coarse maximum has fallen down to the position in the

"tail," and in which only the maximum in the fragments can be clearly traced (samples 60009.454 and 60009.458 [17]). Attempts to identify two constituents in the composition of these samples leads to very high values of the coefficient χ^2/N .

The book by H. L. Alling [11] represented a major contribution to our understanding of the mechanism underlying the formation of a bimodal granulometric composition resulting from the breakage of rock. Alling describes how the initial unimodal distributions of rock-forming minerals resulting from breakage on the initial stage forms an additional maximum in the region of fine particles. The fine maximum then becomes the basic maximum, while the initial distribution gradually decreases, passing through a "tail" stage in the process of falling down to zero. In the course of the grinding process, both maxima shift in the direction of the finer particles, the fine maximum increasing, and the coarse maximum decreasing. Unfortunately, Alling used only graphs to illustrate the study, and numerical data are not presented, which means that we cannot apply our technique, and which, therefore, precludes any quantitative comparison of the experimental data with the natural formations.

In summarizing our discussion, we must emphasize that the bimodal nature of the granulometric composition of sedimentary formations represents a reflection not of any specific conditions of sedimentation, but of a more general physical process, that of brittle breakage of rocks and minerals. The technique that we have employed allows us to divide these types of distributions into constituents of fragments and mineral grains and to estimate the parameters of each. A bimodal granulometric composition is, as a rule (here we are speaking of a probabilistic characteristic), inherent to lunar regolith, moraine, and cryogenic residual rock, i.e., to those types of deposits in the development of which brittle breakage has played a major role. Under the sorts of natural conditions in which these types of deposits form, the probability for particles to shatter depends not on the dimensions of the particles, a factor that was investigated in the models of A. N. Kolmogorov [3] and A. F. Filipov [7], but also on the time and on the degree of consummation of the breakage process. As products of the breakage process, i.e., fine particles, accumulate, the probability that large mineral fragments will break-up drops. By taking this feature into account, we are able to suggest a mathematical model for the formation of the bimodal distribution of particles by dimension as an intermediate product of the process of brittle breakage of rock. While it is true that the end result of this process is difficult to achieve even over geological times, some kind of unimodal distribution will be reached in accordance with any one of the mathematical schemes that have been previously proposed [3, 7, 12].

The superposition of other types of natural processes on top of the process of brittle breakage of rock leads to distortion of the bimodal distribution of particles by dimension. The technique that we have applied to handle the formation of the solifluction deposits of Khibina made it possible not only to disclose these types of transformations of bimodal distributions, but to also trace their trend. With intensification of the transport processes, dispersal of the fine particles is observed, which initially leads to a growth in the gradation, and then to an increase in the modal dimension of the constituent of the fragments, and, finally to the formation of a unimodal distribution, which, in terms of form, is now quite far from a granulometric distribution of the products of the process of brittle breakage of rocks under natural conditions.

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TECHNIQUES

ESTIMATING PALEOTEMPERTURES AND THE DYNAMIC KATAGENETIC FACTOR OF SEDIMENTARY DEPOSITS FOR THE CASE OF ULTRADEEP WELLS IN CISCAUCASIA

É. S. Sianisyan and A. N. Reznikov

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The rheological and geophysical investigations that have been carried out over recent years have persuasively demonstrated the block structure of the basement as well as the sedimentary complexes of Ciscaucasia. This structure is responsible for the differences in the development conditions and the changes in the rock, fluids, and organic matter, which, in turn, exerts a substantial influence on the volume of oil and gas present in different regions. In order to estimate the producing potential of a deposit it is necessary to have some knowledge of the quantitative criteria of transformation of matter. In addition to the heating capacity of rock, which may be estimated on the basis of the maximal paleotemperature, there is the relative index of dynamic katagenesis, a characteristic that was proposed by A. N. Reznikov in 1988. Theoretical grounds for introduction of the concept along with a quantitative estimation of the relative index of dynamic katagenesis were set forth in [5, 6]. The motivation for the concept derived from the contemporary notion that the dynamics of tectonic processes may play a highly substantial role in the evolution of oil and gas basins at the point when there occurs basically a reduction of the energy barriers against chemical conversions and mass transport, and dynamic tectonic excitation of the rock-fluid system is achieved under the influence either of a shock wave, shearing strain, seismic vibration, variation of the electromagnetic field, and dilatation.

The basic assumption that has to be made in determining the relative index of dynamic katagenesis is to assume that the geothermal factor, which is expressed as a compound index, as an *exponential geochronotherm*, and the dynamic tectonic index, which is provisionally estimated through study of the history of the geological development of the eastern Donets coal basin, represents the principal influence on the reflectivity of vitrinite in the coal inclusions. It has been established empirically [16] that

$$DK = 1.4 \left(\frac{0.9R_0}{\varepsilon_r} \right)^{1.5},$$

where ε_r is the exponential geochronotherm and R_0 is the reflectivity of vitrinite, expressed as a percentage.

Let us consider how these factors exert their effect using as an example the Cenozoic deposits found in ultra-deep wells located within the West Kubin trough. This trough has been characterized by marked tectonic activity throughout its geological development. Prolonged intensive downwarping in the Mesocenozoic resulted in the build-up of a thick sedimentary mantle that reaches 12-14 km in the central portion of the trough, in particular, 6 km and more in the Cenozoic section [2, 3].

The Levkinsk ultra-deep well No. 250 is situated in the Levkinsk anticlinal zone along the southern folded flange of the West Kuban trough, where a significant proportion of the proved oil and gas reserves are concentrated [1]. The brachy-anticline is composed not only of Upper Oligocene levels, but also of Eocene-Mesozoic deposits. The formation of the contemporary state of the rock and fluids within this region is the result of the influence of the Akhtyr sutural zone and a series of submeridional faults that cross this zone.

The northern New Dmitri uplift represents a typical geosyncline contortion fold complicated with upthrow fault types of dislocations with amplitude up to dozens of meters. It is situated in the underthrust portion of the section along the southern flange of the West Kuban trough and belongs to the Levkinsk antisynclinal zone, which is separated from the Kaluga upthrust, the latter having an amplitude 600-1500 m.

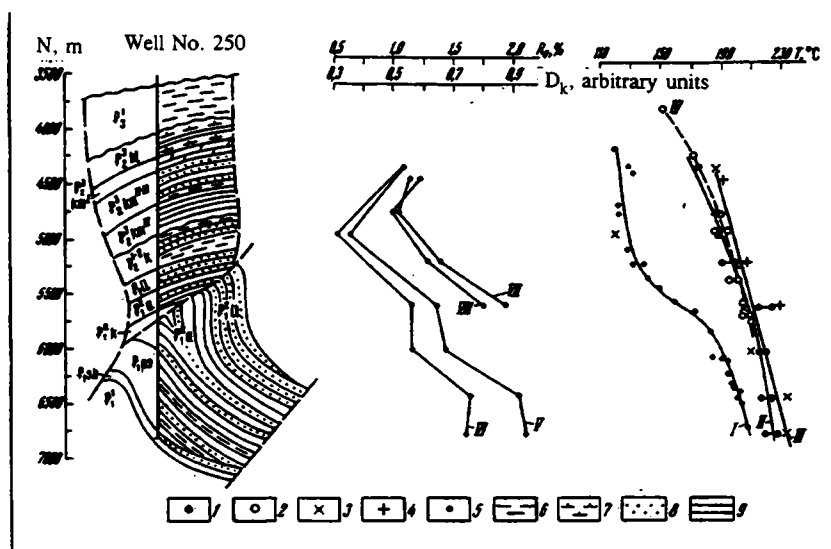


Fig. 1. Paleotemperature and dynamic katagenetic sections of ultradeep wells in the western Kuban trough. (1) and (2) values of the paleotemperatures determined by means of the thermobarogeochemical method, with (1) referring to well No. 250, and (2) to well No. 2; (3) and (4) the same data gathered on the basis of a determination of the reflectivity of vitrinite, with (3) referring to well No. 250 and (4) to well No. 2; (5) contemporary temperatures from well No. 250; (6) clays; (7) clay-marl; (8) sandstone; (9) claystone. I, Thermographs of well No. 250; II and III, paleothermographs determined on the basis of data of thermobarogeochemistry and the reflectivity of vitrinite in well No. 250; IV, paleothermograph determined on the basis of data of thermobarogeochemistry in well No. 2; V and VI, curves showing the variation of the reflectivity of vitrinite and the relative index of dynamic katagenesis in well No. 250; VII and VIII, the same data for well No. 2.

The data that have been obtained from the study of the Cenozoic deposits of the West Kuban trough in the area of the Levkinsk and northern New Dmitri structures and comparisons of these data with the results gained from studies of Mesozoic and Paleozoic platform complexes that have experienced different development histories in terms of tectonic activity are characteristic elements in any general investigation of dynamic katagenetic phenomena and correlation of paleotemperatures that have been determined by means of methods of thermobarogeochemistry and vitrinite thermometry. The results of these tests are presented in Table 1 and in Fig. 1. The reflectivity of vitrinite has been studied at the All-Russia Research Institute of the Oil and Coal Prospecting Institute and the values that have been obtained were recalculated for paleotemperatures by means of a scale developed by the Institute of Geology and Exploitation of Fuel Minerals [4].

Over 30 samples were subjected to this type of analysis. In only six of the samples, however (one from well No. 250, four from well No. 2 from the northern New Dmitri uplift, and one from well No. 4 from the northern New Dmitri uplift) was organic matter discovered on the basis of which the reflectivity of vitrinite could be estimated. For purposes of comparison four samples from well No. 1 (Druzhba), which is situated in the platform portion of Ciscaucasia, were investigated. Organic matter was represented mainly by vitrinite attritus with grain dimension in the range 10-20 m. A total of 15-20 measurements of the reflectivity of vitrinite per sample were performed. Table 1 gives the arithmetic mean for each sample. The thermobarogeochemical determinations were performed by means of a VD-6 vacuum decryptometer using a technique set forth in [7].

The distribution of the maximal temperatures determined by this technique matches the general geothermal law of heating of sedimentary basins and is close to a parabolic law. However, a special feature was found on the paleothermograph of the Levkinsk well associated with a sharp variation of the paleogeothermal gradient in the period from the Upper to the Middle Eocene over the range of occurrence of the Paleogene deposits. Thus, whereas the paleogeothermal gradient within the Upper and Middle Eocene (Kuma and Kutai suites) amounts to 35-38°C/km, in the Paleocene (hot spring formation) it is in the range 5-6°C/km.

TABLE 1. Chronobarothermal and Dynamic Katagenetic Characteristics of Deposits

Well	Depth, m	Age, millions of years	T, °C	R ₀	PT _{RV}	PT _{TB}	D _k	ε _r
250	4361—4366	39	136*	1,08	185	175	0,54	1,64
	4880—4887	40	150*	0,62	120	187	0,31	1,74
	5193—5200	45	—	—	—	190—200	—	—
(Levkin)	5605—5610	57	170	1,36	205	215—225	0,57	1,99
	5992—5997	60	180	1,43	210	215—240	0,57	2,09
	6389—6394	64	200	2,06	235	218—225	0,77	2,30
	6742—6750	65	210	2,11	235	220—230	0,74	2,40
	—	—	—	—	—	—	—	—
2 (North New Dmitri)	3789—3797	31	—	—	—	152	—	—
	4230—4237	32	—	—	—	172	—	—
	4442—4450	33	149	1,16	190	—	0,58	1,68
	4744—4750	38	158	1,08	185	188—190	0,50	1,78
	4881—4890	—	—	—	—	185—195	—	—
	5192—5200	41	170	1,39	207	202	0,62	1,88
	5360—5365	—	—	—	—	195—200	—	—
	5584—5586	50	178	1,87	230	205	0,79	2,0
	5654—5663	—	—	—	—	205—210	—	—
	5735—5739	—	—	—	—	210	—	—
4 (North New Dmitri)	5161—5165	47	—	2,08	235	180	0,91	1,95
1 (Druzhba)	5330—5331	290	175	2,26	240	—	0,70	2,70
	5183—5189	250	170	2,00	235	200	0,55	3,01
	4735—4737	115	115	1,1	187	183	0,40	2,38
	4373—4375	110	150	0,99	173	170	0,39	2,02

*Extrapolated values of the temperatures disregarding convective heat transport. PT_{TB} and PT_{RV} are the paleotemperatures determined by means of methods of thermobarogeochemistry and the reflectivity of vitrinite, respectively.

This great a variation of the geothermal gradient could be associated with a discontinuity, however, judging from the composition as well as the occurrence and development conditions [1, 2], it was not that significant and could not account for the paleotemperature discrepancy. In that case it is also difficult to explain the sharp drop in the paleotemperature gradient in deposits of the hot spring formation. The cause, we believe, could be activation of tectonic movements in the western Kuban trough. The Akhtyr overthrust, which may be followed along the southern part of the Levkinsk uplift, stimulated the expression of different forms of shoves, predominantly of the upthrust type, as a result of which tectonic dislocations are found in the Neogene, Paleogene, and, it would appear, more ancient complexes. Analysis of core material extracted from the ultradeep Levkinsk well No. 250 demonstrates that in the range of depths 5500–6700 m of rock, represented by argillites and aleurolites, which in some sections are deformed, friction planes are established and thin hairline cracks with calcite filling may be found, while within the rock mass of the Kipyachyi ravine (5700 m) analysis of the geological and geophysical data show that all the indicators of an upthrust fault are present.

Thus, the decrease in the paleogeothermal gradient, which is confined to deposits from the hot spring formation, is attributable primarily to processes induced by compressive strain as a result of the Akhtyr overthrust (cf. Fig. 1). The elevated values of the relative index of dynamic katagenesis (cf. Table 1), which lead us to assign the southern flange of the western Kuban trough to intensively excited geological structures, underscore the high dynamic tectonic activity of the region within the wells we have studied.

Calculated values of contemporary temperatures were used to estimate the relative index of dynamic katagenesis for sedimentary complexes of the Levkinsk well.

Regression analysis of the dependence of the relative index of dynamic katagenesis (DK) on depth yielded the following equation: for well No. 250, $DK = 0.190H - 0.46 + 0.18DK$ ($r = 0.79$), for well No. 2, $DK = 0.194H - 0.34 + 0.10DK$ ($r = 0.81$), where H is the depth of occurrence of the sample, km, and r is the correlation coefficient.

With increasing depth and, correspondingly, a growth in the age of the rocks, the value of the relative index of dynamic katagenesis grows as a linear function, a result which is confirmed by comparison of the values of the relative index of dynamic katagenesis for samples extracted from the same ranges of depths from two districts, but belonging to

different stratigraphic complexes (cf. Fig. 1). However, detailed analysis of individual values leads us to a number of regularities.

In the overthrust portion of the Levkinsk structure (range 2.8-5.4 km) the gradient of the relative index of dynamic katagenesis is only 0.037 1/km, while in the underthrust (5.4-6.7 km) it is five times greater (0.190 1/km). In the geological column (see Fig. 1) it is clear that the strata in the underthrust are markedly shifted and deformed. It is evident that, besides tangential forces, additional strains directed across and diagonal to the stratification have been created here under the influence of the tectonic factor. That is, these compressive strains were caused, we believe, by a growth in the gradients of the relative index of dynamic katagenesis and of the reflectivity of vitrinite. In the more submerged zone, i.e., at greater distances from the upthrust plane, the rate of increase of these indices falls. In the column from well No. 2 at depths in the range 4.4-5.6 km, the gradient of the relative index of dynamic katagenesis reaches 0.194 1/km. The minimal values of the relative index of dynamic katagenesis in the area of the Levkinsk structure fall in the disconformity region, which is confined to the bottom of the Kuma formation.

Data for well No. 1 (Druzhba), which is situated in the platform portion of the Tersko-Caspian trough, are presented in Table 1 for the purpose of comparing the dynamic katagenetic characteristics of geological columns from different tectonic zones. In the case of deposits at depths in the range 4280-4730 m, which is comparable to the depths in the western Kuban trough, but are more ancient, the value of the relative index of dynamic katagenesis is at most 0.4. The elevated values of the relative index of dynamic katagenesis are confined to the Permo-Trias complex, in which cinerous material attesting to the potential influence of additional superposed phenomena associated with volcanic activity has been found.

Another consequence of stress compression of deposits, in addition to the dependence on the age of the deposits, is the entry into the interstitial space of water from closed cavities in the rocks and minerals, along with activation of processes of dehydration and dilatation. The release into the intergranular space of new portions of water helped to facilitate the movement of laminated rock (argillites), which, in turn, induced active redistribution of heat as a result of convective mass transport. Such a hypothesis is confirmed by the maximal temperatures that have been established in samples extracted from depths in the range 5592-5997 m, where decryptive activity up to 240°C has been noted. Another confirmation for initiation of heat redistribution in certain sections, possibly through overflows, is the correlation of the paleotemperatures and contemporary temperatures in the well, the values of which largely converge with increasing depth. Whereas at a depth of 4365 m within the Kuma formation this difference amounts to 55°C, in deposits of the hot spring formation at a depth mark of 6750 m it is only 10-15°C. The averaged thermographs also have the same shape. However, the inflection in the thermograph curve that was found is situated somewhat higher (at depths of 5500-5600 m, whereas in the contemporary thermograph it is at depths of 5900-6100 m).

Such a temperature distribution corresponds to a geological column in which a disconformity between the occurrence of the Paleocene deposits of the Kipyachyi and Akhtyr ravines has been established at depths of 5700-5750 m as a result of an upthrow fault. The maximal paleotemperatures determined by the thermobarogeochemical method and reconstructed over this range of depths exceeds the values of the averaged paleothermograph by 20-10°C, a result that may be attributed either to local effects due to friction between the rocks or convective heat transport in lateral migration of fluids into autochthonous strata delivered subvertical to the upthrust plane.

Deflections of the paleothermographs and of contemporary thermographs coincide with the depth of the upthrust plane, due to the intensive convective heat and mass transport both in the thermally stressed as well as in the contemporary stage of development of the geological structure. The facts that we have presented give us grounds for assuming that, by comparison with the present, intensive redistribution of heat in the past (including redistribution by means of convection) extended over a broader range of depths in the region of tectonic dislocations.

Data from estimation of the paleotemperatures on the basis of determinations of the reflectivity of vitrinite using the scale developed by the Institute of Geology and Exploitation of Fuel Minerals are worth noting. The values of the paleotemperatures also increase with depth, though the variance is higher than when our own determinations are employed. In view of the fact that quite different methods are used to estimate the maximal heating capacity, the results that we have obtained should be considered entirely comparable. At the same time, there are differences in the temperature indicators that basically tend to overstate the results that are obtained using the reflectivity of vitrinite. As in one previous study [6], we attribute this result to the influence of the dynamic katagenetic phenomena on the vitrinite reflectivity readings, which are not taken into account in the scale developed by the Institute of Geology and Exploitation of Fuel Minerals.

Thus, the results that we have presented from paleogeothermal and dynamic katagenetic investigations carried out in ultradeep wells in Ciscaucasia yield new estimates of the geological history and role of geothermal and dynamic tectonic processes in the past, arrived at on the basis of a new approach. This, in turn, discloses new horizons for more well-founded forecasts for the search for minerals.

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